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Boris S. Sokolov

Academician, Russian Academy of Sciences
Counsellor of the Presidium of the Russian Academy of Sciences

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Vladimir Vasil'evich Menner (On His 90th Birthday)

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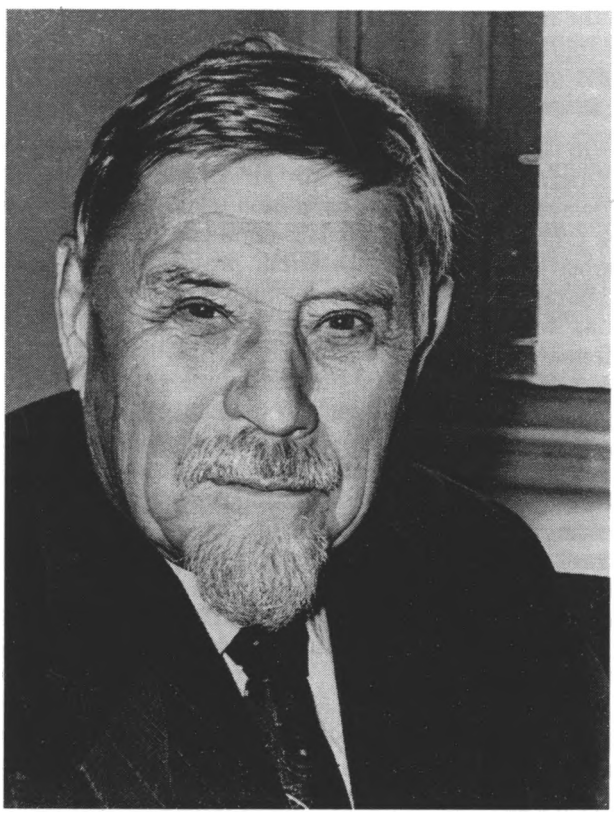

November 24th is the 90th birthday of Academician Vladimir Vasil'evich Menner, a prominent Russian geologist and expert in stratigraphy and paleontology, who devoted more than 70 years of his life to national geology. Two volumes of his *Selected Papers* were published in 1991–1992, and at present, a book of recollections about Menner is under development. *Selected Papers* (vol. 1, General Problems of Stratigraphy; vol. 2, Regional Stratigraphy) manifested the great role of Menner in advancing geoscience, primarily stratigraphy, in our country and abroad in recent decades. In the book of memoirs, Menner is recalled as a marvelous person who taught his disciples to be persistent in pursuing their goals and chosen orientations of scientific work, develop foresight, and search for new pathways in science. His devotion to his beliefs, sense of justice, kindness, and human compassion have earned him the respect and fascination of all who knew him. Menner will always be remembered as a distinguished representative of the Russian intelligences. His name is associated not only with certain contemporary scientific trends in stratigraphy: he was also known for remarkable organizing activity, he educated and trained three generations of paleontologists and stratigraphers in our country, and had a long and very distinguished record of involvement in many international scientific organizations. For these reasons, we would like to commemorate his bright and versatile scientific, pedagogical, and organizational career, as well as his charming and unique personality.

Menner was born on November 24, 1905, in the town of Shatsk. When he was 17, he entered the division of natural sciences of the Physical-Mathematical Faculty at Moscow State University, and he graduated in 1927. His post-graduate studies were done under the supervision of A.P. Pavlov, when he was a scientific assistant in the Mining Academy. Just three years after receiving his diploma, Menner was appointed Acting Head of the Chair of Paleontology of the Moscow Geological Prospecting Institute, and was further appointed as its Head until 1966. In 1934, Menner was awarded degrees in biology and geology without defending a thesis.

In 1934, Menner began to work at the Geological Institute (GIN) of the USSR Academy of Sciences, and was closely associated with this institute for the rest of his life. He was both Head of the Laboratory and Sector of Phanerozoic Stratigraphy and Deputy Director of the Institute.

It was at this institute that his scientific and creative abilities were most conspicuously manifested. The Sector of Stratigraphy at the Institute, which had long been active and successful, was Menner's creation. The scientists of the Laboratory, with Menner at the head, succeeded in solving a number of broad and vital stratigraphic problems, which actually embraced the entire stratigraphic scale, from the Precambrian to the Quaternary.

Even in the 1940s–1950s, while dealing with problems of regional and applied geology, Menner frequently referred to general problems of stratigraphy. His doc-

torate thesis (1961), devoted to biostratigraphic methods of correlating different sedimentary facies, significantly contributed to geoscience. In the 1950s, Menner was among the founders of the Interdepartmental Stratigraphy Committee of the USSR, where he served as Vice Chairman for 35 years. The great role of this Committee in fostering stratigraphic research in our country is widely recognized.

In recognition of his outstanding scientific achievements, Menner was elected Academician (in 1966) of the USSR Academy of Sciences. In 1973, Menner became Vice Chairman and then, from 1978 until the end of his life, Chairman of the National Committee of Geologists of the Soviet Union (and then Russia). He was also among the creators of the International Geological Correlation Program (IGCP), and served as its Vice Chairman. For many years (from 1972), Menner was editor-in-chief of the journal *Izvestiya Akademii Nauk SSSR, Seriya Geologicheskaya*, a member of the Editorial Board of the journals *Vestnik Akademii Nauk SSSR* and *Doklady Akademii Nauk SSSR*, a member of the Presidium of the Moscow Society of Naturalists (MOIP), a member of the Technical Board of the Ministry of Geology and Environment Protection of the USSR, and from 1966, Head of the Department of Paleontology at Moscow State University. At the 27th session of the International Geological Congress (IGC), held in 1984 in Moscow, Menner was Deputy Chairman of the Organizing Committee and Chairman of the Scientific Program Committee.

Menner's influence in science was confirmed by the highly authoritative professional and honorary positions that he held. He was honored to be President (from 1968 to 1972) of the Commission on Stratigraphy of the International Union of Geological Sciences (IUGS), President of the International Subcommission on Paleogene Stratigraphy (1972–1976), Vice-President and (since 1976) member of the IUGS Executive Committee, a member of the French, London, and Indian Geological Societies, a member of the New Zealand Royal Philosophy Society, an honorary member of the Canadian Society of Petroleum Geologists, and a member of the Academy of Sciences of Hungary. For his great services to national geoscience, Menner also received various governmental awards.

Behind the above, seemingly bare facts and dates, lies the bright scientific activities of a talented, keen investigator and remarkable organizer, who educated several generations of stratigraphers. Scientific treatises by Menner are a valuable contribution to national and international science, as well as to theoretical and practical stratigraphy. He is the author of nearly 250 scientific works.

In the first period of his scientific career (the 1920s–1940s), Menner dealt with paleontological problems of Mesozoic–Cenozoic vertebrates and their use in stratigraphy. Particularly interesting was his research on Upper Mesozoic and Paleogene ichthyofaunas: they provided an early clue to the detailed correlations of Upper Cretaceous and Paleogene deposits in southeastern oil-bearing provinces of the European part of the USSR.

Menner's stratigraphic works contributed much to advancing this branch of science in our country. In the 1920s–1940s, his attention was also focused on problems of regional stratigraphy, such as the Mesozoic–Cenozoic areas of the Crimea-Caucasian region, Paleozoic deposits of Baku II, and ancient sequences of the Polar Urals. In each of his endeavors, Menner brought with him extensive expertise to increase the knowledge of the studied regions. In each case, he attempted to use stratigraphic data to solve complicated and vital problems of general geology and tectonics in theory and practice.

In his subsequent activity, he viewed regional stratigraphy as the basis for all stratigraphic studies. However, in the 1940s he turned to issues concerning the criteria for the subdivision and correlation of sedimentary sequences in different provinces, and to the elaboration of stratigraphic taxonomy. These investigations were interrupted by World War II, and when Menner returned to them in the 1950s, he was able to meet the challenges of practical geology of that time. The incipient average-scale geological survey required essentially new correlation techniques and had to operate with a uniform stratigraphic scale of the USSR in order to elaborate a basis for the prognosis and prospecting of mineral resources. The methodical aspects of correlation and the general stratigraphic scale for the territory of the USSR were the main concern for Menner for many years. In addition, he directed the division (later sector) of stratigraphy. Menner's monograph *Biostratigraphic Basis for Correlating Marine, Lagoon, and Continental Formations,* published in 1962, assimilated the worldwide experience which had been accumulated up to that time and became a significant event for developing new insights into stratigraphy.

The guidelines of the stratigraphic nomenclature and classification, which were worked out in the 1950s–1970s with Menner's active participation, are also noteworthy. They enabled us to produce more precise regional scales, acquire a unified terminology for legends of geological maps, and finally, find a basis for elaborating stratigraphic codes for the USSR.

By applying the methods of regional and interregional correlations to noncoeval sequences in the USSR and other countries, Menner had the opportunity to elaborate on general stratigraphic problems and to refine the principles and structure of the general stratigraphic scale.

It would not be an exaggeration to say that Menner's works created a coherent conception of the modern stratigraphic scale, substantiated its subdivisions, and outlined the methods for its further development and detalization. Menner was persistent in supporting the

complex substantiation of subdivisions of the general scale.

All of Menner's works illustrate that he was always seeking for general regularities against a given regional background. He proposed the now well-known hierarchic system of stratigraphic subdivisions, which clearly separated three principal parts of the scale: Precambrian, Phanerozoic, and Quaternary. The chronological span of these subdivisions (from hundreds and dozens of Ma up to hundreds of ka) is chiefly determined by the different methods used to identify them (historical–geological, biostratigraphic, or climatostratigraphic). These particular ideas of Menner were the subject of a special discussion at the 27th session of the International Geological Congress in Moscow in 1984.

Another important stratigraphic problem to which Menner made a significant contribution is the detalization of stratigraphic scales, using new groups of organic remains and new techniques. From the 1960s, despite the sceptical and ironic attitude of many prominent researchers, he actively supported the subglobal character of both stages and zones. During the last years of his life, Menner dealt with the problem of the wide utilization of infrazonal (microstratigraphic) units.

Emphasizing the reality of subdivisions of the general stratigraphic scale, in the 1960s to early 1970s Menner turned his efforts toward clarifying the stages of development of marine basins and the organic world as well as investigating the problem of changing paleogeographic environments. Without a doubt, many of his concepts resulted in the appearance and development of a prospective branch of recent stratigraphy, which is frequently referred to at present as ecostratigraphy or eventual stratigraphy.

Menner's activity as an organizer of science was invaluable. It was he who transformed the Geological Institute of the USSR Academy of Sciences into a leading institution in stratigraphy gathering numerous geologists, stratigraphers, and paleontologists, and educating a professionally strong scientific team. The Geological Institute is the *alma mater* of many doctoral and candidate students, for which the approbation by stratigraphers of GIN was a guarantee of their high scientific level.

All colleagues, followers, and acquaintances of Menner respected and admired him. The enormity and diversity of his knowledge and his excellent memory were remarkable. Moreover, he approached each problem with a lively and keen interest, which allowed him to deeply understand any given problem under consideration, evaluate all of its aspects, and identify priorities. His sincere devotion to science helped him to search and find ways, sometimes untraditional, to settle problems at hand. Owing to these human features, Menner was always the center of attention. There can hardly be a stratigrapher or paleontologist in our country who had not turned to Menner for advice and received his assistance and encouragement. Thanks to his encyclopedic knowledge and his ability to see the general through details, Menner could, with equal expertise, discuss stratigraphic problems of the Riphean and Paleogene, the taxonomy of foraminifers and acritarchs, or the stratigraphic application of conodonts and stromatolites.

Menner took up new problems with great enthusiasm. Thus, he became involved in the development of new biostratigraphic methods and study of nontraditional fossil groups. Such groups were once foraminifers, later palynologic remains, and in recent times, conodonts and Paleozoic radiolarians. Through Menner's efforts, these groups were introduced into practical geology. Among his most significant initiatives over the last 25–30 years was the organization of Precambrian biostratigraphic research on the basis of studying acritarchs and stromatolites. The progress made by our investigators in the study of Precambrian biostratigraphy stimulated similar activities abroad. Ergo, a great deal of credit for the success in Precambrian biostratigraphy, where our scientists are among the leaders, must go to Menner.

It is noteworthy that although he was a biostratigrapher, Menner was also interested in many other methods of stratigraphy. He was among the first to evaluate the significance of paleomagnetic data in stratigraphy, he favored the utilization of paleoclimatologic methods, in particular, for exploring Precambrian tilloids, and contributed to elaborating the balanced use of the radiometric and seismostratigraphic methods.

Menner was a man of strict principles and was persistent in pursuing his scientific goals. At the same time, however, he was tolerant of his scientific opponents. He was always ready to consider and discuss controversial issues and to evaluate the proposed arguments. While he sharply criticized slight and insufficiently substantiated opinions, his criticism was always just and beneficial, and his comments were interesting and fruitful. Along with criticism, Menner's statements also contained many positive aspects, and everybody was eager to hear his opinion on a presented report or thesis. Many times, Menner suggested absolutely new approaches to resolving seemingly trivial problems. These approaches were further evolved in publications, and in some cases, served as bases for dissertations.

Menner was always willing to share his ideas and considerations on how to approach particular problems or issues. These human features united a large team of researchers around him, who presently continue to work and solve problems successfully.

One of the most attractive features about Menner as a leader was the complete absence of any formality in his behavior, and his readiness to examine any problem in detail. He was also able to find common features in two or more competing conceptions, and unite seemingly inconsistent viewpoints. In many cases, it was useful to find the best solutions for both scientific and organization problems.

Menner highly evaluated the scientific achievements of other researchers, and clearly expressed this in relevant articles. The greatest esteem he had for his educators A.P. Pavlov, M.V. Pavlova, A.D. Arkhangel'skii, and N.S. Schatsky, and he devoted several wonderful articles to them.

Menner was unique as a man and as a scientist; he infected his colleagues and students with his overwhelming energy, which was manifested in all his affairs, major and minor.

Menner belonged to a category of scientists, rare today. He combined the encyclopedic knowledge typical of all prominent scientists of the 20th century with deep insight into the different problems under study, characteristic of scientists with a narrow specialization. Thus, it is quite natural that he contributed so much to geoscience in this country, and had numerous disciples and followers who now work in various fields of geology. Foreign scientists consider him to be one of the most outstanding representatives of the world scientific community.

Menner was a man of the highest moral standards. He inherently combined high professionalism with extraordinary human qualities. We shall remember him as an inspiring leader, unfailing colleague, and a fascinating person. Once more rendering our respects to Menner, we recall the words of a poet: "we ought not utter with a heavy heart, Alas, he is no longer with us." We should bless the stars that we had been his colleagues and associates.

The ideas of Menner, an outstanding scientist of our time, are a valuable contribution to modern geology, particularly, to stratigraphy. His creative heritage will certainly serve as a basis for new scientific discoveries for many years to come.

Reviewer B.S. Sokolov

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The School of Thought of Academician V.V. Menner

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The school of thought of V.V. Menner should be the subject of a profound investigation by historians of science. But in connection with his birthday, one cannot but touch upon this topic, as the scientific school he created is just one of the major results of his activity, as well as a reflection of his unique personality. This school cannot be attributed to a certain institution, city, or country, as Menner's disciples work in different cities and countries. This school of thought was formed as a consequence of his intellect, enthusiasm, moral example, and charm. Therefore, it is appropriately called the school of Menner.

Academician Menner did not personify the commonly accepted image of those prominent in science and social consciousness: a purposeful lone investigator who left voluminous collected works devoted to specific fields of science and who had followers, but not necessarily disciples. He was unlike those administrators of science who almost displaced such lone investigators. One can rarely find significant personal works or basic ideas of "organizers" in scientific reviews or in libraries. They are surrounded by coauthors but not by disciples. Menner published a great deal of original work and led many projects, but, in my opinion, the large school of thought is his greatest legacy.

In the late 1940s and early 1950s, newcomers to stratigraphy were struck by Menner's authority in the field, as he was not yet a candidate of science. He was always at the center of stratigraphic meetings, discussions, scientific councils, excursions, and seminars. He was constantly surrounded by people who wanted to talk, consult, or argue with him, or to publish, organize, or receive something. Memories of the lines of people wanting to meet with him by his office in the Geological Institute are still vivid. Among those surrounding him there were not collaborators of the Department of Stratigraphy from the Geological Institute and other subordinates, but rather, visitors. The impression arose that he was one of the centers of Soviet stratigraphy. Newcomers understood, with time, that Menner, a Candidate of Biology and Geology, and later Doctor and Academician, was in fact such a center. Further, he was not an administrative, but rather, an intellectual authority, who originated a great deal of basic ideas and projects, which were subsequently developed and disseminated throughout the country by numerous researchers, postgraduates, and doctorates. This is how Menner's school of thought was formed.

It is not easy to outline the boundaries of any school of thought. It is especially difficult to do it for the school of Menner. It was an extremely open association. Menner actively involved researchers from different institutions and scientific schools in the execution of his projects and programs, and numerous scientists willingly took part in those investigations. Menner's school of thought, therefore, did not have distinct formal boundaries, and it was difficult to evaluate its real scale and influence, though they were obviously great. An attempt to gain some idea of its impact was made in 1985 in connection with Menner's 80th birthday, when researchers of the Geological and Paleontological Institutes and Moscow State University intended to compile a bibliographic handbook of monographs published by scientists who considered themselves to belong to the school of Menner. It was apparent that this could only be accomplished with the participation of those who had actually worked under his leadership, or had been among his postgraduate students. However, this idea was flawed, as it could not include all the disciples of Menner, namely, all of his former students and researchers who had since deceased. In addition, it was not easy to establish which scientists belonged to Menner's school, as some mature researchers also readily joined groups, including informal ones, led by Menner. At the same time, a number of his disciples passed to other institutions, where they brought his ideas and methods and developed some new ones. As a result, very interesting investigations proceeded, namely, those of Repina *et al.* (1964), Alekseeva (1967), and others. Nevertheless, the bibliographic handbook which resulted did set forth some objective idea of the school of Menner. The handbook, conventionally entitled *Collected Works of the Menner School of Thought,* was produced as a single copy in three bound albums and presented to the hero of the anniversary with the autographs of all the monograph authors. The handbook contained covers, title pages, and contents of monographs arranged by theme.

Looking through this handbook now, one can see that its significance goes far beyond that of an anniversary present; it is an interesting document on the history of geology and a valuable source of information, first, on the scientific activity of Menner and, second, on the

history of geology in the USSR and Russia in the middle and late 20th century.[1] It is also of methodological interest for scientific schools. There are apparently no similar publications, at least in the field of geology. Therefore, at the end of the paper, the handbook under consideration will be included, edited and represented as a list of works.

This list, in contrast to the handbook, represents monographs arranged in alphabetical order and not by theme. Additionally, it is supplemented with a number of authors and works which were previously dropped from the book. However, not all the gaps are closed. The list notoriously understates the scale and influence of Menner's school because it is limited by monographs published prior to 1986, though the school continues to develop. This is evidenced by the fact that in the last decade, the second generation of researchers who belong to the school published significant works (in some cases their most significant ones), and even representatives of the third generation became candidates and published their first monographs. Therefore, it is more accurate to consider the list to represent only the base of the school, as well as its major ideas and areas of investigation as they were formed up to the mid-1980s. Accordingly, we have given it the title "List of the Major Works of the School of Menner."

This list is, in fact, the basis for the conclusions of the proposed paper. Another source is my long-standing work under the leadership of Menner. Beginning from these sources, I will try to characterize, to a first approximation, the school of Menner, with regard to its scale, quality, and themes. The factual evidence of investigations will not be touched upon in view of the limited nature of this paper. As any first attempt, this one certainly has flaws, if not serious disadvantages. However, in my opinion, the occasion justifies its appearance.

The list of major works of the school of Menner yields 185 monographs from 93 authors. It is a whole library in itself! A major portion of the monographs represents the published candidate (57) and doctoral (26) dissertations developed under the leadership of Menner or his elder disciples and collaborators (namely, V.A. Vakhrameev, B.M. Keller, D.M. Rauser-Chernousova, E.D. Zaklinskaya, K.V. Nikiforova, and S.N. Naumova), as well as under the guidance of the next generation of his disciples (namely, A.S. Alekseev, I.S. Barskov, Yu.B. Gladenkov, S.V. Meyen, L.N. Repina, A.Yu. Rozanov, Kh.S. Rozman, M.A. Semikhatov, V.V. Khomentovskii, and others). These figures reflect a great stratigraphic school consisting of the three generations of researchers, and known for a great deal of famous names, extensive studies, and ramified lines of investigation. The proportion of collective works and collections of articles in the list is also very significant (60), as are the monographs on major scientific issues (63) associated with the formers according to their themes. The school of Menner may therefore be represented as a single community, working out several major and urgent stratigraphic problems in cooperation with researchers from other regions, institutions, and scientific schools. Such close cooperation and joint research resulted in a number of well-known books, namely, that of Zhuravleva, *et al.* on Siberian archaeocyathids (1964), *Mezozoiskie vysshie sporovye rasteniya SSSR* (*Spravochnoe rukovodstvo*) by Vakhrameev *et al.* (1980), many papers of Rauser-Chernousova with coauthors, and others.

From the content of the handbook and titles of the monographs, it follows that the theoretical problems of stratigraphy and biostratigraphy, new methods and comprehensive research in stratigraphy, zonal and infrazonal stratigraphy, paleobiogeography, eventual stratigraphy, general stratigraphic scales and principles of their development, as well as regional stratigraphy, were and still are the major focuses of investigation by the school of Menner. These problems were elaborated more or less successively, usually by the example of several stratigraphic intervals, from the Upper Precambrian to Cenozoic, and of different regions, from Iceland to Kamchatka.

In the foreword to the handbook, the editors wrote:

> The following four common features of the school of Menner are the most important. The first is an extreme devotion to science. The second is the exacting requirements of the quality of factual data. The third is a striving to see the general in the particular. The fourth is a joint investigation, together with free individual research.

These words, written by the "pupils" of Menner, well represent the quests, inner atmosphere, and spirit of the school of Menner.

The atmosphere of the school resulted from the unique personality of Menner, whose main features were similarly characterized by most authors of the memoir volume prepared for publication.[2] He possessed all of the prerequisites of a leader of a large scientific school as if he was born for this purpose. And certainly, he could not help creating a school around him. At the risking of repeating myself, I will list those features of Menner, which in my opinion, enabled him to create a large school and favorable atmosphere necessary for fostering the creative work of outstanding scientists during three generations.

Among those features are, first, a great intellect, enthusiasm, energy, and optimism, which aroused in researchers a wish and power for solving scientific problems. The combination of these features imparted a significant dynamism and, I would say, even swiftness to the activity of Menner. His second virtue was great scientific and human intuition, which enabled him to choose promising subjects for research, methods of investigation, and teams of researchers. In deciding on

[1] The three volumes of this reference book are kept in the family of A.K. Bystrova, who kindly placed them at my disposal.

[2] *Vladimir V. Menner as a Scientist and a Person*, Yu.B. Gladenkov, Ed.

subjects and candidates, Menner was rarely mistaken. However, when this happened, he generally did not blame a diligent investigator for the failure, but rather, suggested another theme more suitable for the investigator's character and interests. In most cases, mistakes were almost absent, as Menner closely observed the advances and failures of young scientists during their first attempts, and precisely evaluated their possibilities and biases. He attached particular significance to the latter. This was a manifestation of his regard for people, especially those who were diligent and interested in science. In deciding on a new theme, researchers usually agreed with the suggestions of Menner, convinced of his arguments and authority, if this was not the case, he did not insist. However, when he intuitively felt that he made the proper choice of a promising theme and investigator, he resorted not only to persuasion, but also to power. Many examples confirmed the force of his intuition, including rather intricate cases. It is known that he induced Vakhrameev to investigate Mesozoic flora, Semikhatov to study stromatolites, and me to deal with tillites (after my failure with endemic stromatolites from the Patom uplift).

The ethical climate in the school of Menner was marked by high morality, correctness, and responsibility, all of which he himself exemplified, and this contributed to the unity of the disciples around him. Individuals who disturbed this climate did not stay for too long in the school of Menner. Menner generously shared ideas with his disciples, but unlike many others, he never took credit as a coauthor. In fact, he even refused any suggestion of this kind. Menner appreciated interests of science, geology, and stratigraphy higher than any group or personal interest, be it his own or that of certain persons in the Stratigraphic Department, Geological Institute, Academy of Sciences, etc., and he taught us the same. Once in my youth, when I told Menner that I had to go urgently to do field work in the Lena gold-bearing region to forestall collaborators from the All-Union Research Institute of Geology and Yakut Geological Survey with the correlation of Upper Precambrian sections, he answered unusually sharply: "It makes no difference who will do it, but it should be done well." There were many similar episodes, and these were also lessons in the school of Menner.

Not only the businesslike efficiency of Menner, but also his significant virtue played an important role in the rise of his school of thought. He attracted people, and kept them within his orbit with the stability of the school. Neither his subordinates nor colleagues saw or felt his command in the ordinary sense. His charm, democratism, benevolence, and ability to behave and discuss problems with us on equal terms as if he was simply the elder colleague, aroused the wish to ask advice, discuss issues, and confide in him with our successes and failures. When Menner criticized someone, both in public and confidentially, he did it very benevolently with the wish to help. He never restricted himself to criticism, and always analyzed the reasons of failures and suggested possible ways of improvement. Menner also remained good natured while listening to critical remarks about himself even if he vigorously rejected them. Thus, for the most part, his disciples are not afraid to be criticized and to criticize themselves, considering this to be a normal and indispensable part of the scientific process.

Menner thoughtfully trained his younger colleagues, both in the Geological Institute, and in other institutions; he kept a close watch on their successes, and tried to help, where possible, either practically or with advice and interesting information. He often initiated this help himself. For example, he summoned a certain researcher and told him: "You should attend such-and-such all-union (or international) conference; it is very important for you. I have already included your mission in the plan and I will press for it." He would also, for example, unexpectedly approach a junior researcher and quickly tell him about interesting facts and ideas extracted from a report or publication which would be of use to the person. Other times, encountering someone in the corridor, he would enthusiastically say: "I have listened to your communication (or "I was told about your report"). It is very interesting. Be sure to write a paper on this subject. We will publish it at once in the *Izvestiya Akademii Nauk*"; or "You have collected intriguing data, but the conclusions are not sufficiently convincing. What about trying to process them this way?" In his impetuous life charged with events, he possessed a strikingly good memory. He also thought about things important to his disciples, knew how we were getting on, and what we needed for success in science and in life. He, himself, often thought of different ways to help people, rather than doing it only in response to our request.

Vladimir V. Menner is a remarkable trace in our souls, just as his school of thought is in Russian science.

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The Regional, Theoretical and General Stratigraphy as Treated by the Interdepartmental Stratigraphic Committee (ISC)

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Abstract—The 40 years of activity of the Interdepartmental Stratigraphic Committee (ISC) of Russia for the benefit of the regional stratigraphy in the country are briefly reviewed; the most significant results of the work of the ISC and its Commissions on the problems of theoretical and general stratigraphy are summarized for the first time. The estimated completeness of the stratigraphic studies is supplemented by a formulation of the key problems of regional stratigraphy that are common for many regions and require special research. The theoretical research in stratigraphy has included the object, subject and basic principles of science, problems of typization and classification, principles used in terminology, and criteria for the determination and correlation of the stratigraphic units and their boundaries. The results of this work were published in the two editions of *Stratigraficheskii kodeks* (Stratigraphic Code) issued in 1977 and 1992. The general stratigraphy covers the problems of unification of terminology and nomenclature, preparation of stratigraphic scales and schemes, unification of stratigraphic procedures, and elaboration of methods. Recommendations are given for theoretical research and for the study of the general problems of stratigraphy.

Key words: *stratigraphy, regional, theoretical, general; Interdepartmental Stratigraphic Committee, scale, scheme*

The present Statute of the Interdepartmental Stratigraphic Committee (ISC) was ratified on June 16, 1992, by the Committee on Geology and Exploitation of Mineral Resources of the Russian Federation (*Roskomnedra*) and the Department of Geology, Geophysics, Geochemistry and Mining Sciences of the Russian Academy of Sciences. The first paragraph of the Statute declares:

"The Interdepartmental Stratigraphic Committee (ISC) of Russia is the supreme interdepartmental body that organizes and coordinates the stratigraphical service in Russia, provides scientific and methodological administration of stratigraphic research on the territory of the country and in individual regions, and considers and issues resolutions on urgent problems in stratigraphy." (*Postanovleniya Mezhvedomstvennogo ...*, 1992, p. 7).

The ISC was set up in 1955 to work out the stratigraphical principles and provide the required conditions for obtaining research data for the compilation of the State Geological Map of the USSR, scale 1 : 200000.[1] In the course of its 40-year history, the ISC was constantly considering the growing requirements of practical geology and the acute necessity of organizing theoretical, general, and methodological studies in stratigraphy.

Important improvements were introduced in practical stratigraphy, such as the establishment, in 1965, of the Regional Interdepartmental Stratigraphic Commissions (RISC), the transition from compilation of general regional stratigraphic schemes to those of local areas, the application of the results of zonal stratigraphy, the study of reference sections, and other enterprises.

In terms of methodology, the biostratigraphic and geochronologic methods were supplemented by climatostratigraphic, magnetostratigraphic, and seismostratigraphic, e.g., by setting up the commissions on Magnetostratigraphy and on Shelf Stratigraphy. Considerable efforts were made to promote the application of complex methods.

The problems of theoretical and general stratigraphy were studied by practically all the commissions, but primarily by the commissions on Phanerozoic Systems, on Precambrian Subdivisions and on Stratigraphic Classification, Terminology, and Nomenclature (CSCTN). In 1994, the VSEGEI convened an interdepartmental meeting on the general problems of stratigraphy (Zhamoida and Prozorovskaya, 1995; *Postanovlenie Mezhvedomstvennogo ...*, 1994).

The same tendency characterizes the participation of the ISC in international stratigraphic organizations, such as the International Commission on Stratigraphy (ICS), its subcommissions and working groups. The ISC also took part in the scientific meetings of other associations.

[1] The resolution to inaugurate the Committee was adopted in 1955 at the All-Union Meeting on the General Problems of Stratigraphic Classification convened at VSEGEI. A booklet on stratigraphic and geochronologic units (*Stratigraficheskie i geokhronologicheskie podrazdeleniya*, 1954) was issued for the public discussion of practical and theoretical problems in stratigraphy.

V.V. Menner, during his chairmanship of the ICS from 1968 to 1972, updated the structure of the Commission by using the relevant experience of the ISC (Menner, 1974).

The publishing activity was an important aspect of the work of the ISC. During its four decades, the ISC published about 45 issues of Resolutions of the Interdepartmental Regional Stratigraphic Conferences (IRSC), complete with the schemes for the USSR territory, 22 volumes of *Trudy* (Collection of Works), 27 issues of *Postanovleniya* (Resolutions) of the ICS itself, two editions of the Stratigraphic Code, and 11 booklets on methodology. Actually, the Commissions on Systems prepared the state stratigraphic correlation schemes, which were published in the volumes of *Stratigrafiya SSSR* (Stratigraphy of the USSR), *Geologicheskoe stroenie SSSR* (Geological Structure of the USSR), *Geologiya SSSR* (Geology of the USSR), and others.

The manifold activity of the ISC is briefly described in several articles (Vereshchagin and Zhamoida, 1977; Zhamoida, 1968; Zhamoida and Prozorovskaya, 1985, 1989, 1995; Menner, 1982; Sokolov, 1987; *Stratigraphicheskaya klassifikatsiya*, 1980; and others) and in the issues of ISC Resolutions, including those dedicated to the results of the work of the Commissions (*Postanovleniya Mezhdvedomstvennogo* ..., 1976, 1978).

Academician D.V. Nalivkin was the first Chairman of the Committee and has served as its Honorary Chairman since 1976 (Zhamoida *et al.*, 1983). His closest assistants were N.K. Ovechkin, Vice-Chairman (until 1962), Academician V.V. Menner, Vice-Chairman (until 1989), and N.N. Bobkova, Scientific Secretary (until 1978). We now list the first Chairmen of the commissions on Precambrian subdivisions (S.V. Obruchev, Corresponding Member of the Academy of Sciences of the USSR, and Academician N.S. Schatsky [soon followed by B.M. Keller]), and on Phanerozoic systems (T.N. Spizharskii, B.S. Sokolov, D.V. Nalivkin, A.P. Rotai, D.L. Stepanov, B.K. Likharev, L.D. Kiparisova, G.Ya. Krymgol'ts, V.P. Rengarten, I.A. Korobkov, A.G. Eberzin, and E.V. Shantser). From 1976, Academician Sokolov was the Chairman of the Committee (since 1988 he has been the Honorary Chairman). A.I. Zhamoida, Corresponding Member of the Russian Academy of Sciences, has been the Chairman of ISC since 1988 and, at present, the Vice-Chairmen are Academician M.A. Semikhatov, A.Yu. Rozanov, V.A. Prozorovskii, and Yu.B. Gladenkov; E.L. Prozorovskaya has been the Scientific Secretary since 1976.

The general ISC structure includes the Bureau of the Committee, the members of the Committee (*Postanovleniya Mezhvedomstvennogo* ..., 1992), and three groups of Commissions: for regional coverage, subdivisions of the general stratigraphic scale, and research subject matter.

The Regional Interdepartmental Stratigraphic Commissions (RISC) are set up for the northwestern regions (V.I. Robonen, Chairman), the central and southern parts of the Russian Platform (S.M. Shik), Ciscaucasia (Yu.Ya. Potapenko), the Urals (B.I. Chuvashov), Siberia (F.G. Gurari), and the Far East (M.T. Turbin); there are also groups for the northeastern regions (K.V. Simakov) and Sakhalin–Kamchatka area (B.A. Sal'nikov).

The commissions for the Precambrian units and the Phanerozoic systems are those of the lower Precambrian (V.Z. Negrutsa), upper Precambrian (M.A. Semikhatov), Cambrian (A.Yu. Rozanov), Ordovician and Silurian (T.N. Koren'), Devonian (M.A. Rzhonsnitskaya), Carboniferous (A.Kh. Kagarmanov), Permian (G.V. Kotlyar), Triassic (A.N. Oleinikov), Jurassic (K.O. Rostovtsev), Cretaceous (V.A. Prozorovskii), Paleogene (V.I. Yarkin), Neogene (Yu.B. Gladenkov), and Quaternary (B.A. Borisov).

The subject matter of the research is under the supervision of the commissions for Stratigraphic Classification, Terminology, and Nomenclature (A.I. Zhamoida), Regional Stratigraphic Schemes (V.I. Yarkin), Reference Sections (N.N. Predtechenskii), Magnetostratigraphy (A.N. Khramov), Stratigraphy of the Shelfs (I.S. Gramberg), and Isotope Geochronology (A.G. Rublev).

Several ISC Regional Commissions of the former Soviet Union have subsequently become national stratigraphic bodies, and are now Associated Members of the ISC of Russia (*Postanovleniya Mezhvedomstvennogo* ..., 1994). The Baltic Stratigraphic Association, set up in 1990 on the basis of the Baltic RISC, maintains close contacts with the ISC.

During the last forty years, the ISC has issued systematic reviews of its regional work, whereas the scope of the activity of the Committee and of its commissions in the field of theory and general stratigraphy has never been surveyed. In the present paper, we try to fill this gap by using the abundant materials published in the ISC resolutions and in other issues. We admit the arbitrary division of the spheres of application of theoretical and of general stratigraphy, however, the sequence of representation that we have assumed allows us to show the work of the Committee from many aspects, and to formulate recommendations for the future.

We start our review with regional stratigraphy, which is closely related to the organization and improvement of the stratigraphic service in the country.

REGIONAL STRATIGRAPHY

As the main task of the ISC management, the results of stratigraphic research in the different regions of the country were repeatedly published and generalized in the resolutions of ISC meetings, in the regulations of the Committee and corresponding two volumes of *Stratigrafiya SSSR* (The Stratigraphy of the USSR), in the two issues of *Geologicheskoe stroenie SSSR* (The Geologic Structure of the USSR), in the seven volumes of *Stratigraficheskii slovar'* (The Stratigraphic Dictionary), and in a number of special review papers. We

show our estimations of the results of the latest interdepartmental stratigraphic meetings (up to 1995) by appending a table with traditional representation to this article.

Herein, we intend to discuss the main problems of regional stratigraphy, which are of common importance for many regions and areas of Russia. Their solution can be reached by specific research to meet the requirements of the legends for the series of *Gosgeolkarta-200* (The State Geological Map-200), which constitute the main task of the complex geological mapping of the country.

(1) The primary problem is still the definition of subdivisions, their age, and the correlation of continental rocks. Corresponding recommendations were adopted by the IRSC concerning the upper Precambrian (red-colored rocks) of North European Russia, the Silurian of Tuva, Tatarian deposits in the Urals, Jurassic of the Urals and western Siberia, Cretaceous of western and middle Siberia, and of the northeastern regions. The only approach to the solution of these important problems is the application of complex methods.

(2) The correlation of continental and marine coeval deposits is still to be solved for the Devonian of the Kuznetsk Alatau, Gornaya Shoriya, and Primor'e; for the Permian of the Pechora basin, Taimyr, Verkhoyansk region and southern Primor'e; for the Triassic of the Caspian region and Polar Urals; for the Induan Stage of the middle part of Siberia; for the Jurassic of the Angara–Vilyui depression; and for the Paleogene of the eastern slope of the Urals. Again, the main causes of these gaps in our knowledge are an incomplete set of research methods, a disregard for the time markers, and insufficient data on microfauna and palynoflora.

(3) Several meetings were convened in the last few years on the Precambrian highly metamorphosed formations, though in practice, every region still has numerous unsolved problems. Therefore, the tables showing the extent of Precambrian research, in most cases, qualify the regional and local schemes as working scales. Consequently, this reflects the considerable decline in the number of geochronometric laboratories, lack of geochemical studies, and drastic reduction of paleontological studies on the Precambrian, especially on microfossils.

(4) Among the most dramatic situations is the study on the age, subdivision and correlation of volcanogenic, volcanogenic-terrigenous and volcanogenic-siliceous rocks. The Silurian and Devonian of the Urals, Triassic of western Siberia, Ordovician of middle Siberia, and corresponding formations of nearly all ages in eastern Russia belong to this group of rocks. In recent years, a number of utterly unexplored problems appeared in connection with the condensed sections of siliceous rocks (e.g., from Permian to Jurassic in the Far East), and with the so-called chaotic complexes. The solution can be found by studying conodonts and radiolarians; this approach has already been tested in several regions, e.g., in the Far East and in the Urals.

(5) We shall mention only two of the numerous problems of Quaternary stratigraphy. One of them is the comparison of polygenetic deposits of glacial and ice-free areas, where the glacial, i.e., climatostratigraphic concept, competes with the morpholithostratigraphic approach. Another problem concerns the correlation of Anthropogene on land and in shelf areas. Both problems are particularly acute for western Siberia and parts of the Far East.

(6) Besides the priority of the biostratigraphic method in improving the stratigraphic basis of the Phanerozoic, we should emphasize that the insufficient elaboration of local and provincial biozonation scales for different floral and faunal groups is often the main cause of the unreliability of stratigraphic schemes. Owing to the scarcity of funds, paleontological research in this country has almost ceased, and we should do all we can to help the researchers who are still involved in these studies and attract young scientists.

(7) The resolutions of every regional meeting include recommendations for refining the positions of stage (not to mention the series) and even systems boundaries in the local sections. A low precision in this respect greatly hinders the mutual adjustment of *Gosgeolkarta* sheets and especially of their series. It is obvious that these problems can only be solved by applying paleontological methods.

(8) Among the various physical methods, the isotopic–geochronologic (radiometric) ones are of particular importance. Moreover, the foremost task is the establishment of a new geologic time scale of the Phanerozoic in Russia. After G.D. Afanas'ev's well-known scale, issued in 1964 and 1975, these studies were practically abandoned. A close analysis of the foreign scales demonstrated that even their latest versions are based on datings, most of which cannot be assumed as markers. This work should be undertaken by VSEGEI, the Geological Institute of RAS (GIN), the Institute of Geology of Ore Deposits, Petrography, Mineralogy, Geochemistry (IGEM), and the Institute of Mineralogy, Geochemistry and Crystallochemistry of Rare Elements (IMGRE). The efforts of these institutions should also include the restoration of the network of corresponding laboratories in the regions.

(9) A specific and, from many points of view, complicated problem is setting up the state computer data bank(s) for all regional and local stratigraphic correlation schemes, with facilities for their operative correction. The *Stratigraficheskii slovar'* (The Stratigraphic Dictionary) of Russia should be used as one of the blocks of the data bank, continually revised and supplemented. It is the task of the principal institutes on stratigraphy of the *Roskomnedra* (the Russian Committee on Mineral Resources) and the Russian Academy of Sciences, in cooperation with the ISC, to carry out this work.

Results of Interdepartmental Regional Stratigraphic Meetings according to ISC records

Eonothem	Region / System	Russia								Adjacent territories					
		Russian platform	Ciscaucasia	Urals	Western Siberia	Middle Siberia	Northeast	Sakhalin, Kamchatka	Far East and Transbaikal region	Baltic region	Belarus	Caucasus	Kazakhstan	Middle Asia	
										The year of the last meeting of IRSS					
										1976	1981	1955, 1958	1986	1958	1971
Cenozoic	Quaternary	U, O (1983, 1984)		U	U (1988)	U (1979)	U, C, O (1982)	C, O (1982)	U, O (1982)	U	U	U, C	C	C	
	Neogene			O (1990)	U (1976)	C	O (1975), U (1989–1992)	U, C, O (1989, (1992)	O	O	C	U, O	U, C	O	
	Paleogene			C	U	C			C	U	O	U, O	U, C	O	
Mesozoic	Cretaceous	U, O (1990)	U (1978)	U	U, C	U, C (1978)	O	U (1974–1975)	C, O	U	U		U, C, O		U, O
	Jurassic	U (1988)	U (1977)	U (1977)	U, C (1990)	U, C	U		U, O	U	U		U		U, O
	Triassic	U, C (1979)	C (1977)	C	C	U, C	U		U	O	O		U, C, O		U, O
Paleozoic	Permian	U		U, O		U, C, O	U, C		U, O	U	O		C, O	O	
	Carboniferous	U (1988)		U		U, C	U		U, C, O (1990)	O	C		U, C	O	
	Devonian	U		U		U, C, O (1979)	U, O (1974–1975)	O (1995)	U, C, O	U	U		C, O	O	
	Silurian	U (1984)		U (1990)		U, C, O	U		O	U	U		U, O	U	
	Ordovician	U, O		U		U, C			O	U	U		U	O	
	Cambrian	U (1983)		C, O		C, O (1986)	O		C, O	U	U		C	O	
Archean–Proterozoic		O (1974, 1983)		U, O (1990)		U, O (1979), C (1986)	O		U, O	U, O	U, O		O	O	

Note: Scheme types: U – unified scheme; C – correlation scheme; O – scheme under elaboration. The years of the last meetings are shown in brackets.

THEORETICAL STRATIGRAPHY

The most comprehensive treatment of the notion of "theoretical stratigraphy" was accomplished in 1974–1985 in the works of Meyen (1989). The more extensively used term "general stratigraphy" partly includes the theoretical problems proper, to which we refer in this paper, the definition of the object and subject matter of science, its basic principles, typification in stratigraphy, principles of terminology, concepts and structure of stratigraphic classification, relations between different groups and categories of stratigraphic units, criteria for their recognition and correlation, and identification of their boundaries.

The most important discussions on the theoretical problems of stratigraphy were organized by the ISC in the course of preparation and discussion of the drafts of stratigraphic codes and the codes themselves at the meetings of the Commission on Stratigraphic Classification, Terminology and Nomenclature (CSCTN), at the ISC plenary sessions (Kovalevskii, 1971, 1980; *Postanovleniya Mezhvedomstvennogo* ..., 1969, 1973, 1977, 1992) and at the meetings of its commissions, or in publications. It was not happenstance that the first volume of ISC works was dedicated to the analytical review of the codes adopted in other countries (Zhamoida *et al.*, 1969).

Among the stratigraphers whose theoretical research had a significant influence on the concept and principles of the Russian codes and on many important resolutions in these codes, we should first mention A.N. Krishtofovich, D.L. Stepanov, B.M. Keller, L.S. Librovich, A.P. Rotai, V.V. Menner, B.S. Sokolov, and S.V. Meyen; extensive work was carried out by the compilers of the codes and many other active participants in the discussions, the results of which were included in the ISC Resolutions and in the prefaces to the codes.

Below is a list of fundamental theoretical concepts which were used as a basis for the Russian codes and worked out as a result of discussions over the course of many years.

(1) The codes do not contain the definition of stratigraphy. The Introduction to the second edition of the code, however, contains an ample exposition of its concept formulated by Meyen. But in the book by Meyen (1989), in the Russian codes, and in practical work, the sphere of stratigraphic studies includes the geologic bodies, which are subjected to the so-called law of superposition. These geologic bodies, i.e., the stratigraphic units, are actually objects of science. The purposes of stratigraphic research are the identification of the units, determination of their space–time relations (correlation), and improvement of the general (global) stratigraphic scale.

(2) There are four basic principles of stratigraphy; namely, the objective reality and uniqueness of stratigraphic units (Stepanov and Mesezhnikov, 1979), Steno's principle, Huxley's principle, and Meyen's principle (Meyen, 1989; *Prakticheskaya Stratigrafiya*, 1984).

(3) Proceeding from Steno's principle, the preeminence of stratigraphy proper is acknowledged; i.e., the temporal relations of geologic bodies are determined by their spatial relations.

(4) The unity of stratigraphy as the fundamental branch of geological sciences is recognized in contrast to the concept assuming the possibility of numerous "stratigraphies," which correspond to different methods and are ultimately independent.

(5) In stratigraphy, typification is an important process. Librovich and Ovechkin (1963) worked out the principles of determination and the rules for selection of stratotypes and of their variants. This was not only the basis for the Russian codes, supplemented by the notion of a limitotype, but also for international manuals on stratigraphy and some national codes.

(6) Many discussions were dedicated to determining the principles of stratigraphic classification and, consequently, of terminology (the latter is described below in *General Stratigraphy*). As a result, the principle of the so-called unified stratigraphic scale was rejected, though it was the essential idea in the booklet *Stratigraficheskaya klassifikatsiya i terminologiya* (1956). Moreover, the system of multiple classification, adopted in the American Code and in the international manuals issued in 1976 and 1994, was declined.

The stratigraphic units are divided into two groups: the basic and specific. The latter, though qualified as independent, is supplementary to the basic. The geographic criterion is regarded to be of crucial importance for the classification of basic units divisible into three categories: general (global), regional, and local. The geographic criterion is also essential in the classification of certain categories of specific units, like the paleontological principle when applied to classify the basic units of the Phanerozoic. Specific units (litho-, bio-, climato-, magneto- and seismostratigraphic) are established by a variety of methods.

The stratigraphic units of different groups and categories are independent.

(7) For the first time in stratigraphic practice, the general stratigraphic taxonomic scale was officially extended by the introduction of a higher taxon, i.e., the acrothem, and of the lower taxa, i.e., the division, link, and step. The concept and introduction of a horizon and a provincial zone (lone) were new additions to stratigraphic classification. The elaboration of the notion of a formation can also be regarded as a considerable achievement, despite current criticism.

(8) The interrelations between the categories of local, regional, and general units are assumed as the relation of the particular to the general (Teslenko, 1976). The general units have a potential planetary distribution and are established by means of combining various methods of study according to the position and

the rank of a given subdivision in the geological section of the crust.

(9) The criteria for identifying and ranking the basic subdivisions is their possible (at a given state of study) correlation with the stages of development of all geospheres for the general units, or with the stages of geological history of particular crustal areas for regional and local subdivisions. This approach emphasizes the geosystem character of the basic taxa.

The introduction of the concept of lateral boundaries for local and regional units (after Krishtofovich) was an important progressive step.

Within the rather wide range of theoretical research of the ISC and its commissions, the study of certain problems connected with geological practice should be continued. The most important among these problems are time and stratigraphy, the principal possibilities and restrictions of the rhythmo-, climato-, chemo- and seismostratigraphic methods, elaboration of sequence stratigraphy, the principles of improvement and detalization of the general stratigraphic scale for the Precambrian, and the status of biozonal subdivisions of different categories and their correlations. A special elaboration of criteria is necessary for the determination of stratigraphic boundaries of subdivisions of different categories and ranks with due account of the significance of hiatuses in sedimentation.

GENERAL STRATIGRAPHY

We include into the list of problems of general stratigraphy the unification of terminology and nomenclature, elaboration of stratigraphic scales and schemes (the practical part of this problem concerns the regional stratigraphy), the procedure of identifying general, regional (provincial) and local stratigraphic subdivisions, and a unified approach to other stratigraphic procedures. The work on these problems actually involves the preparation of stratigraphic codes. We also include the development of various methods into the scope of general stratigraphy.

With regard to terminology, it was traditional to adopt the resolutions of the International Geological Congresses (units of the general scale), follow the traditions of Russian geological practice (regional and local units), and suggest the optimal number of new terms. As a result, the geological practice officially admitted only about 25 new terms: straton, acron and acrothem, lone, phase, division, link, step, climatolith, stadial, holo-, lecto-, neo-, hypo- and parastratotypes, limitotype, areal stratotype, the names of magnetozones, and some other terms. We consider the limited number of new formal terms as a positive factor; this measure was later confirmed in geological practice with the rejection of dozens of new terms suggested by authors, some of whom are of considerable repute. In this case, we should be guided by the principle of necessity and sufficiency.

The work on the problems of nomenclature was carried out according to traditions of geological research in Russia (Zhamoida *et al.*, 1968; Krishtofovich, 1945; Moiseeva, 1980; Khomizuri, 1960). First of all, the name construction for a stratigraphic unit was clearly defined: "The full stratigraphic name should consist of two parts: the stratigraphic term (the nomenclature term—the word to be defined), indicating the taxonomic rank and category of a stratigraphic subdivision, and the name proper (the word defining the term)" (*Stratigraficheskii kodeks*, 1992, p. 61).

For the first time in Russian literature, the concepts of the unit's validity, its authorship (and coauthorship), and priority were formulated in addition to basic principles of the stratigraphic nomenclature, such as the approval, renaming and preservation of unit names, double names of units, synonymy and homonymy, and so on. Both issues of the Russian Stratigraphic Code include "The Rules of Formulation and Spelling of the Names of Stratigraphic Subdivisions."

No other national stratigraphic code or international manual has such a scrupulously elaborated nomenclature along with a convenient exposition of it for geological practice as the Russian codes.

For example, in the literature published abroad, the stratigraphic scale or chart are practically the same as the scheme. While the same practice exists in our country too, the codes proposed to make a distinction between them, and the corresponding definitions are now used in practice (*Stratigraficheskii kodeks*, 1977, 1992).

The ranking of the stratigraphic correlation schemes in accordance with the reliability of their substantiation (unified, correlative, and operative) was an important innovation aimed at a more objective assessment of the state of the stratigraphic basis for concrete areas and individual regions.

The identification of geological systems, series, and stages as the most settled general units and the more accurate definition of the boundaries between them are tasks that the ISC discusses only periodically, in most cases, following the resolutions of international forums. This is actually how the Ordovician System and its subdivision into three series (1975) was established, the boundary between the middle and upper Jurassic was drawn at the top of the Callovian (1989), and the Cretaceous–Paleogene boundary over the top of Maestrichtian (1984) and the separate Berriasian Stage were acknowledged (1965). The discussions about the volume of the Quaternary System, which was being resolved for many years, were settled by adopting the resolutions of the International Commission on Stratigraphy (ICS) on lowering the boundary of the system down to 1.6 Ma (1990) and approving the system as a separate unit (1959, 1961, and 1987).

In the meantime, Russian geologists submitted their own proposals on the improvement of the general stratigraphic scale. In this respect, the ISC has adopted a number of important resolutions that were accepted

or are gradually being accepted in world practice. The foremost is the substantiation of the individuality of the Paleogene and Neogene systems (1956). Only in the semiofficial *Global Stratigraphic Chart of 1989* (Cowie and Basset, 1989) was the necessity of this subdivision of the Tertiary system acknowledged.

Among others, a principally new idea was to reject the subdivision of the Quaternary System into traditional series and stages, because its entire range corresponds to the zonal subdivision, and the whole duration of the Quaternary Period complies with a phase. The essential contributions of the ISC to the general scale of the Quaternary System and understanding of the specific character of the initial stage of this period are: the four units of the Anthropogene distinguished on a climato-stratigraphic basis (1978), which were later officially termed links (Shantser *et al.*, 1973; *Stratigraficheskii kodeks*, 1977, 1992); the suggestion to subdivide the Pleistocene into Eopleistocene and Neopleistocene, in the division rank of units (1992, 1995); and an attempt to coordinate minor units, such as steps, with the scale of isotopic stages.

The considerable achievements of the corresponding ISC commissions also include the stage and zonal scales of certain systems.

Prior to the 1950s, the Lower and Middle Cambrian was not officially subdivided into stages. In 1956, using the results of studies of the Siberian sections, the ISC adopted a resolution to identify the Aldanian and Lenian stages in the lower series of the system, and the Amgian and Mayan stages in the middle series of the system. As a result of further research in 1982, the scale acquired further details, i.e., the indicated stages of the lower series received the rank of superstages, and were respectively subdivided into the Tommotian, Atdabanian, Botomian, and Toyonian stages (*Postanovleniya Mezhvedomstvennogo* ..., 1983); these stages, similar to the middle Cambrian stages, were internationally acknowledged (Cowie and Basset, 1989). The determination of the upper Cambrian (upsection) Ayusokkan, Sakian, Aksai and Batyrbai stages was also a considerable achievement.

In 1968, the Third International Symposium with field excursions to Podolia and Salair was held in the USSR. At that time, the activity of Soviet stratigraphers and paleontologists was focused on the problem of the Silurian–Devonian boundary, and the Symposium was a good opportunity to initiate the unique international program representing an example of a complex approach to the solution of "boundary conflicts" in stratigraphy. It was also oriented towards elaborating the principles, procedure, and organization of similar research in general.

In addition to the more precise definition of the Lower Devonian boundary, a serious contribution was made by tracing the stages of the Lower Devonian, i.e., Lokhkovian, Pragian, and Emsian, in almost all regions of the Soviet Union, as was recommended by the International Subcommisson on Devonian Stratigraphy. The results practically confirmed the interregional and global rank of these stages. This work was greatly encouraged by the ISC Commission on the Devonian, which worked out a detailed scheme of interregional correlation of Devonian deposits in the main sections of Russia.

In this country, the Carboniferous System is traditionally subdivided into three series, which have an adequate correlation with pertinent subdivisions of the system in other countries. The longstanding dispute regarding the volume and stratigraphic position of the Namurian was solved in 1974, when the Serpukhovian Stage (lower Namurian) was incorporated into the scale to complete the Lower Carboniferous Series. This and the previously established Bashkirian, Moscovian (Middle Carboniferous), Kasimovian, and Gzhelian (Upper Carboniferous) stages are currently used in international practice (*Osnovnye cherty* ..., 1975).

Serious problems concerning the Permian stages were solved in 1964 by introducing three subseries with two stages each, and a fourth subseries with a range identical to that of the Tatarian Stage. The provincial (regional) stage scale was likewise suggested for the Tethyan (Caucasus–Sinian) region. In 1979, the ISC adopted the Permian scale of the Tethyan region (Eastern Paratethys), with the Lower Permian composing the Asselian, Sakmarian, Yakhtashian and Bolorian stages, while the Upper Permian was subdivided into the Kubergandian, Murgabian, Midian, Djulfian and Dorashamian stages (*Ob"yasnitel'naya zapiska* ..., 1980; *Osnovnye cherty* ..., 1984).

On the suggestion of L.D. Kiparisova and Yu.N. Popov (1956), the lower series of the Triassic System, previously identified only with the Scythian Stage, was divided into two stages: the Indian and the Olenekian. The latter two stages are now included in the international nomenclature, though with the assumption that they will eventually be further detailed and possibly raised to the rank of a superstage.

In 1964, the ISC decided to unite the upper and lower Volgian substages of the Jurassic into one Volgian Stage: the equivalent of the Tithonian terminating the Jurassic System of the international scale.

To refine the stage subdivision of the Paleogene, an attempt was made to base them on stratotypes represented by the Bakhchisarai section in the Crimea for the Paleocene–Eocene, and by the section of the Black Sea depression for the Oligocene (1964). Further research, however, showed that eight new stages determinable in these sections are of provincial (regional) importance, and valid for the Eastern Paratethys only. In 1985, the ISC confirmed this resolution of the Commission on Paleogene.

In 1975, one of the ISC commissions suggested a provincial stage scale for the Neogene of the Eastern Paratethys, which was arbitrarily correlated with that of the Mediterranean region. It includes nine stages

(regiostages) for the Miocene and three stages for the Pliocene (*Stratigraficheskii kodeks*, 1992).

There are different viewpoints regarding the status of stages and zones (chronozones). The ISC considers them to be units of the general stratigraphic scale, as it is fixed in both issues of the *Stratigraficheskii kodeks*. The ISC commissions on Phanerozoic systems were constantly busy with working out the zonal schemes that could be included into the general scale. Some of these schemes were admitted, with certain reservations, into almost all of the systems. A significant achievement of Russian stratigraphers is the elaboration of zonal scales for all subdivisions of the Cambrian System. For some of the systems, the ISC had to chose among competing zonal scales based on different faunal groups (e.g. for the Silurian, Carboniferous, partly Cretaceous, and Paleogene), or among "parallel" provincial scales (Permian, partly Triassic and Jurassic, and Paleogene).

Special collective monographs were devoted to zonal (and stage) subdivisions of three Mesozoic systems (*Obshchaya shkala* ..., 1984; *Zony yurskoi* ..., 1982; *Zony melovoi* ..., 1989).

The deficiency of the general Cretaceous zonation, which was recommended for the territory of the Soviet Union in 1981, was soon acknowledged by Russian stratigraphers, principally, it was flawed in terms of the paleobiogeographic peculiarities, which was discussed in the resolutions of several international colloquiums between 1959 and 1983. In 1994, a new variant of the zonal scale for both series of the Cretaceous System was published in *Postanovleniya MSK*. It greatly differs from the scale suggested earlier.

The results of work of the ISC commissions were published in the monograph *Zonal'naya stratigrafiya fanerozoya SSSR* ... (1991), and this issue demonstrates the different perceptions of zonal terminology.

The territory of the USSR provides great opportunities to study nearly the entire range of the Precambrian; therefore, the initiators of several significant modifications of the general scale of the Precambrian were Russian geologists from the ISC Commissions. Until the 1970s, however, only two large Precambrian subdivisions were included into the general scale: the Archean and the Proterozoic (at first, in the rank of group or erathems, later in the rank of eonothems); the position of the boundary between them remained debatable. The Archean and Proterozoic subdivisions were treated as regional, and even local, and their correlation was conventional. Therefore, the corresponding volumes of the *Stratigrafiya SSSR* for the European and Asiatic parts of the country were published separately.

The research, carried out for many years by the ISC commissions on the Precambrian, consummated with the Meeting on Stratigraphy in Ufa in 1977 (*Postanovleniya Mezhvedomstvennogo* ..., 1983). The participants confirmed the official status of the Archean and Proterozoic; the latter was divided into the lower and upper (Riphean), which in turn had four units. The Vendian was accepted as the uppermost unit of the Precambrian.

Further studies with due regard for the world experience showed that the identified units are reasonably correlative, and new data were used to refine the earlier datings of their boundaries and exclude the fourth unit of the Riphean (Kudashian). These were also the cardinal data used for subdividing the Archean into the lower and upper units separated by the boundary with an age of 3150 ± 50 Ma, and for ranking the general units of the Precambrian (1990), which is of principal importance. The Archean and Proterozoic were ranked as acrothems, and their subdivisions (the Karelian and Riphean of the Proterozoic) as eonothems. The Riphean was divided into three erathems, and the Vendian was ranked as a system (*Postanovleniya Mezhvedomstvennogo* ..., 1992; Semikhatov *et al.*, 1991; Sokolov, 1964, 1979; *Stratigraficheskii kodeks*, 1992).

Therefore, the general stratigraphic scales of the Precambrian and Phanerozoic were for the first time "linked" formally. It is essential that general units of the Precambrian were actually discriminated on a stratigraphic, and not a geochronological basis, as it was recommended by the *International Stratigraphic Guide* (1994).

The stratigraphic methods. Many Russian stratigraphers have traditionally considered the basic stratigraphic units (at least regional and local) as units with combined substantiation. The so-called "eventual stratigraphy" has been studied in this country for a long time, perhaps without sufficient attention to the unit boundaries. The organized Commission on Reference Sections (1978) and corresponding instructions for their study (1963, 1983) show that our scientists were ahead of their foreign colleagues in this field of research.

Naturally, the ISC was always interested in new methods of stratigraphic research. The Commission on Shelf Stratigraphy (1978) elucidated various possibilities, and formulated objectives of the seismostratigraphic method and its application in geological practice. The Commission on Magnetostratigraphy (since 1988) considered the principal results of respective studies and outlined trends of their further development. The activity of these commissions and the development of the climatostratigraphic method, under the guidance of the Commission on the Quaternary System, have provided materials to expand the second edition of the Russian Stratigraphic Code with corresponding chapters. The listed methods have found their proper place in the combined studies on the subdivision of sections, identification of stratons, and stratigraphic correlation (both interregional and with the general stratigraphic scale).

The unification of stratigraphic procedures was one of the important tasks of the ISC from the time of its inception. This was necessary for a country with

a vast territory and highly diverse geological structures, and corresponded to general objectives formulated for stratigraphers by the State Geological Survey. The unification presumed universal requirements for the identification and names of stratigraphic units different in rank; for the procedure, presentation and evaluation of stratigraphic research; for the organization of working teams to prepare the materials to be discussed by the ISC; and for the activity of the Committee itself and its commissions.

These problems are condensed in *Stratigraficheskii Kodeks* ... (1977, 1992). The respective instructions are obligatory to the Scientific Editorial Council on Maps of Geological Content and to its branches. They were also included in different handbooks for geological surveys, corresponding textbooks, and *Prakticheskaya stratigrafiya* (1984).

The basic efforts in general stratigraphy, aside from those mentioned above, were concentrated on problems concerning the general stratigraphic scale, such as follows: the detailed subdivision of the Archean and Proterozoic; the stratigraphic relationship between the Riphean and Vendian; the definition of unit ranks in the Ordovician and Silurian, which is used in this country and abroad; the proposal concerning the new stages of the Carboniferous; the number of series in the Permian; the boundary position between the Jurassic and Cretaceous (the problem of the Volgian Stage); the new zonal scale of the Cretaceous; and the validity of the adopted Neogene stages. The systems with abundant continental (including volcanic) formations (primarily, the Permian, Triassic, and Cretaceous) should be further studied to elaborate their regional and interregional stratigraphic scales correlated with zonal scales based on marine organisms. Of particular importance is the selection and description of global stratotype boundary points for general units, which can be presented for adoption by the International Commission on Stratigraphy. The more extensive use of various physical, chemical and other instrumental methods is the requisite condition for progress in stratigraphy.

To conclude, we shall emphasize that the scientific, organizing, and methodological activity of the Interdepartmental Stratigraphic Committee of Russia is focused toward persistent improvement of the stratigraphic basis for all kinds of geological studies and, primarily, for geological mapping, as the primary task for the next decade will be the preparation of the State Geological Map of Russia.

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Dynamics of the Global Diversity of Proterozoic Stromatolites. Article 2: Africa, Australia, North America, and General Synthesis

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Abstract—The dynamics of the global diversity of Proterozoic stromatolites is analyzed at specific and generic levels on the basis of separate records from each of six superregions (the data on three of them are discussed herein), with a subsequent generalizing examination of the superregional taxa samples. The analyzed eleven time intervals of comparable duration include four of the Early Proterozoic and seven of the Late Proterozoic. The samples of the identified stromatolitic taxa from Africa, Australia, and North America are of comparatively low volume. Their diversity (and abundance) peaks are distinguished at 2.0–1.8 Ga ago (Australia, North America), in the Early Burzyanian (Africa, Australia), and in the Early (Africa) or Late Karatavian (Australia, North America). The histogram of the global stromatolite diversity is principally based on the taxa distribution in the largest samples from China and Northern Eurasia. It is of a trimodal character (the peaks at 2.3–2.0 Ga in the Early Burzyanian and Early Karatavian), and demonstrates three periods of low taxonomic diversity at the very beginning (2.5–2.3 Ga) and end (2.0–1.65 Ga) of the Early Proterozoic and during the Late Burzyanian. A sharp diversity decrease is notable in the Late Karatavian (0.85–0.65 Ga), and this trend smoothly extends into the Vendian and, probably, Cambrian time. The secular changes in the Proterozoic stromatolite diversity are depictable against the background of discrete global tectonic events, variations of Sr-isotope composition in the sea water, significant climatic fluctuations, directed changes in pCO_2 and sea water saturation with respect to Ca and Mg carbonates. In addition, during the Late Riphean (Karatavian) and Vendian changes in stromatolite diversity were correlative with secular variations in the volume ratios between the buried carbonate and organic carbon, as well as with some important evolutionary changes in the biota. The dynamics of the global diversity (and abundance) of Proterozoic stromatolites reflects a directed evolution of Proterozoic ecosystems.

Key words: *stromatolites, diversity, ecosystems, Proterozoic, Riphean, Vendian*

Changes in stromatolite diversity and abundance during geological history have drawn the attention of geologists for more than 30 years (see Awramik, 1992; Semikhatov and Raaben, 1993, 1994 for bibliography). At present, both parameters have been substantiated as increasing from the Archean to the Late Proterozoic, first reaching a maximum, and then to be rapidly reducing at the end of the Precambrian and during the Cambrian, thus reflecting a general stromatolite decline at the transition to the Phanerozoic. However, both the character of corresponding histograms (monomodal or more complex) and the precise time of the stromatolite decline met different interpretations (Awramik, 1971, 1992; Walter and Heys, 1985; Cao, 1988; Walter *et al.*, 1992; Grotzinger, 1990; Semikhatov and Raaben, 1993, 1994).

In order to elucidate the diversity dynamics of the Proterozoic stromatolites, we analyzed them at both specific and generic levels with stratigraphic resolution two times greater than before. Another specific feature of our analysis was that the separate study of the records from the six largest superregions—Northern Eurasia, China, India, Africa, Australia, and North America—preceded the analysis of the global diversity based on the summarized superregional samples of taxa. This approach allowed us to minimize the statistic noise related to the nonuniform understanding of the one-rank taxon volume by specialists of different schools, and to better understand the correlation between stromatolite diversity and geological events of different scales. The materials on Northern Eurasia, China, and India have been published (Semikhatov and Raaben, 1993, 1994). This article deals with the records from Africa, Australia, and North America, followed by synthesis of superregional data. It also discusses the possible relations between fluctuations in global stromatolite diversity and other events in the biosphere.

The refined stratigraphic scale adopted for the Precambrian of Northern Eurasia within the former USSR (Semikhatov *et al.*, 1991) served as the time framework for the analysis. For our purpose, we delimitated four informal units within the Lower Proterozoic (Karelian) of this scale (2500 ± 50–1650 ± 50 Ma), which virtually coincides with the Palaeoproterozoic of the geological time scale accepted by the International Subcommission on Precambrian Stratigraphy (Plumb, 1991). These units, with boundaries corresponding to 2300,

2000, and 1800 Ma, are equal or similar in their time range to the Palaeoproterozoic units of the International Subcommission scale. The Riphean–Vendian interval of the Proterozoic (the Meso- and Neoproterozoic) was subdivided into seven units: two in each Riphean erathem (in the Lower, Middle and Upper Riphean) and the Vendian itself. The isotopic ages of the lower boundaries of the Riphean erathems and the Vendian system are known to be 1650 ± 50, 1350 ± 20, 1000 ± 50, and 650 ± 20 Ma[1] (Semikhatov *et al.*, 1991), and the boundaries between the named subordinate units of the Riphean erathems approximately correspond to 1450, 1200, and 850 Ma (Semikhatov and Raaben, 1993, 1994).

Estimating the frequency distribution of stromatolite taxa, we took into account not only data from available literature, but also our own materials as well, including the results of the examination of the type stromatolite collections described by C. Walcott, C.L. Fenton, M.A. Fenton, R. Rezak, S.H. Knight, and J. Bertrand-Sarfati, and the identifications of the Proterozoic buildups collected in North America. The statistic samples of formal species (forms) include those species and varieties described in the literature which we considered valid, and those stromatolites formally identified in publications at the generic (group) level or in the open nomenclature, which were either the only representatives of a group in the corresponding age interval, or possess a specific microstructure. Species cited in literature with question marks were not taken into account in our calculations.

DYNAMICS OF THE PROTEROZOIC STROMATOLITE DIVERSITY IN AFRICA, AUSTRALIA, AND NORTH AMERICA

The presented analysis is the final step in our study on the superregional distribution of stromatolites (Semikhatov and Raaben, 1993, 1994). The information on stromatolites from South America (for bibliography see Trompette, 1994) and the Antarctic is too fragmentary to be considered here.

Africa

Recent isotopic-geochronological and partly chemostratigraphic data corrected the traditional views on the ages of some Precambrian and Cambrian stromatolite-bearing deposits of Africa. At present, the space–time distribution of these deposits seems to be as follows. The Cambrian stromatolite-bearing successions are recorded in the Anti-Atlas and High Atlas Mountains (the northern margin of the continent). The Vendian localities are in the Nama River basin (southwestern Africa), in the Volta depression (Equatorial Africa), and in the Anti-Atlas. The Riphean (mainly the Upper Riphean) stromatolite-bearing deposits of Africa have a wider distribution—they are known in the southwestern (the Damara and Gariep belts), equatorial (the Kibaran, Katanga and West Congo belts, Congo syneclise, etc.), and western parts of the continent (the Gourma basin, Moroccan Anti-Atlas, and especially Taoudeni syneclise, where these deposits constitute bodies unique in extension). The Lower Proterozoic stromatolite-bearing units appear to be absent in Africa. The Transvaal Dolomites of the Kaapvaal craton have long been considered as such a unit, but new isotopic-geochronological data proved its Late Archean age (see below). The Archean stromatolites are known from some other units of the Kaapvaal craton as well, for instance, from the Onverwacht Group (3.54 Ga), the unit enclosing one of two oldest stromatolite localities in the world.

The African stromatolites are poorly studied outside the Taoudeni syneclise. Therefore, the analyzed sample of their taxa (72 species and 39 genera, including nine species and six genera from the Archean) by no means represents the actual diversity, especially since attention was mainly concentrated on columnar buildups and partly on microstromatites, while other morphotypes were not subjected to careful examination.

The taxa subjected to statistical analysis were described or identified by researchers of different schools (Raaben and Zabrodin, 1972; Krylov, 1975; Raaben, 1991; Menchikoff, 1946; Cloud and Semikhatov, 1969; Bertrand-Sarfati and Raaben, 1970; Bertrand-Sarfati, 1972a, 1972b, 1972c, 1976; Bertrand-Sarfati and Eriksson, 1977; Semikhatov, 1977; Schmitt, 1978, 1979a, 1979b; Raaben, 1980; Bertrand-Sarfati and Moussine-Pouchkine, 1985; Bertrand-Sarfati and Vicat, 1987). Nevertheless, most taxa are identified rather uniformly, except for the doubtful identification of some Vendian and Cambrian forms (Schmitt, 1979a, 1979b; Bertrand-Sarfati, 1981). Our knowledge of the stromatolite type collections described by Bertrand-Sarfati confirms that problems concerning the unrecognized synonyms and inadequate understanding of volumes of one-rank taxa are not important with respect to the African material.

As mentioned above, the only Lower Proterozoic stromatolite-bearing unit in Africa was, for a long time considered to be the thick (up to 3.5 km) Transvaal Dolomites. They contain abundant and morphologically diverse stromatolites (Truswell and Eriksson, 1972, 1973), but only some columnar varieties and microstromatites from a limited interval (ca. 80 m) of the only

[1] According to recent opinion, the Varangerian (Laplandian) glaciation marking the beginning of the Vendian, was 610–590 Ma ago and, correspondingly, the age for the lower Vendian boundary should be corrected (e.g., Harland *et al.*, 1990; Knoll and Walter, 1992; Bowring *et al.*, 1993). The opinion is based on the U–Pb age of volcanics from the Avalon zone of North America, where they underlie the Late Precambrian tillites. However, the correlation of these tillites with basal beds of the Laplandian glaciohorizon remains unproved, the more so as the Avalon terrane was a constituent of the Proto-Gondwana of the Late Precambrian time (Rankin, 1993). The isotopic age determinations for rocks of the Laplandian glaciohorizon proper suggest that its lower boundary is ca. 650 ± 20 Ma old (Keller and Semikhatov, 1990; Semikhatov *et al.*, 1991).

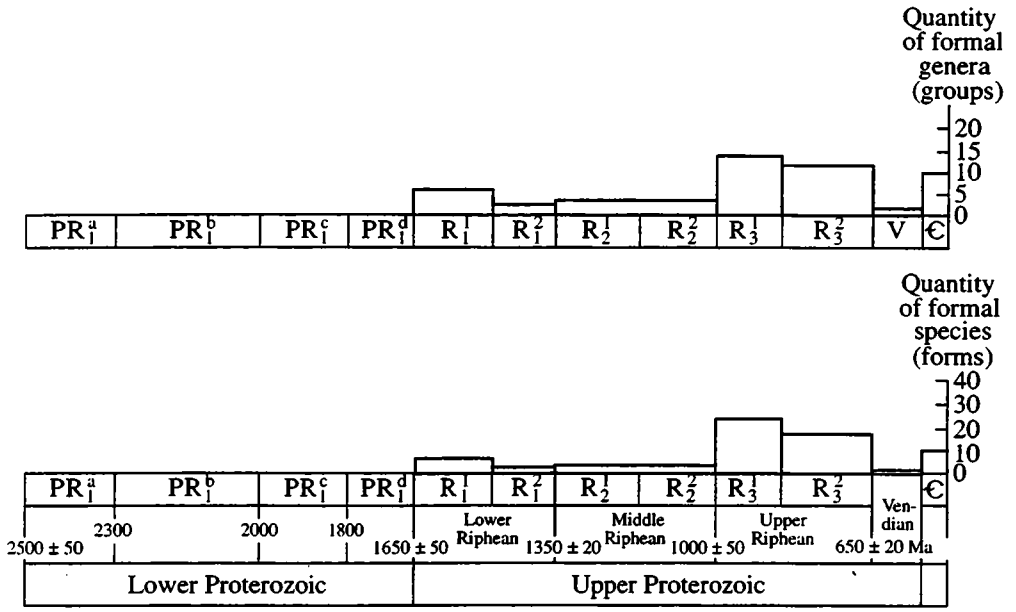

Fig. 1. Secular variations in diversity of Proterozoic stromatolites from Africa at the levels of formal species (forms) and formal genera (groups).

section received a valid description (Bertrand-Sarfati and Eriksson, 1977). Until recently, the age of the Transvaal Dolomites was considered to be about 2.3–2.2 Ga according to Rb–Sr and Pb–Pb isochron datings of under and overlying metavolcanics (Tankard *et al.*, 1982; Walraven *et al.*, 1990). However, the U–Pb zircon age of the Ventersdorp volcanics unconformably underlying the Transvaal Dolomites appeared to be 2709 ± 4 and 2714 ± 8 Ma (Armstrong *et al.*, 1991), and three Late Archean values of isotopic age were measured in the Transvaal Supergroup proper: the isochron Pb–Pb age of 2557 ± 49 Ma for stromatolite dolomite in the middle part of the supergroup (Jahn *et al.*, 1990), and U–Pb zircon dates of 2552 ± 11 and 2520 ± 3 Ma for the volcanics from the uppermost part of the section (Eriksson and Fedo, 1994). These datings prove the Late Archean age of the Transvaal Dolomites. Thus, Early Proterozoic stromatolites seem to be absent in Africa.

The Lower Riphean sections of Africa yielded relatively abundant stromatolites from two units only: from the Kibara Supergroup of the equatorial part of the continent, and from the Precambrian II-2 of the Anti-Atlas (Cahen *et al.*, 1984; Goodwin, 1991). These buildups are paleontologically studied only in the Anti-Atlas, where they form two successive sets of taxa. The first one is dominated by Thyssagetacea buildups (*Cyclopium* sp., *Thesaurus macrotomus* Vlas., *T. pyramoides* Vlas.) and microstromatites with subordinate columnar (*Eucapsiphora*) and nodular morphotypes. Columnar-nodular *Gaia irkuskanica* Kryl. and nodular *Paniscollenia* (Krylov, 1975; Raaben, 1980, 1991) were described in the second set. The composition and succession of the sets suggest that the Precambrian II-2 contains counterparts of the Early and Late Burzyanian assemblages of Northern Eurasia (Raaben, 1991). If these comparisons are correct, the Lower Burzyanian in Anti-Atlas is marked by a greater stromatolite diversity than the Upper Burzyanian (Fig. 1). The scant isotopic-geochronological data do not contradict the assumption that Precambrian II-2 is of the Early Riphean (Burzyanian) age, rather, they show that it is younger than 1700 Ma (Rb–Sr age of the crystalline basement) and older than 680–615 Ma (Rb–Sr age of the Precambrian II-2 metasediments and U–Pb age of intruding granites; Lancelot, 1978; Cahen *et al.*, 1984).

The Middle Riphean of Africa is very poor in stromatolites. The list of reliably identified taxa seems to include only three Middle Riphean forms from carbonate lenses in the siliciclastic Precambrian II-3 deposits of the southwestern Anti-Atlas. These are *Conophyton metula* Kir., *Baicalia prima* Semikh., and *Jacutophyton* sp. (Raaben, 1980). The fossiliferous deposits are cut by granites, which have a U–Pb age of 615 ± 15 Ma (Cahen *et al.*, 1984), unconformably overlay the Precambrian II-2, and are referred either to the Middle Riphean (Choubert and Faure-Muret, 1980; Raaben, 1980; Bozhko, 1984) or to the uppermost horizons of the Upper Precambrian (Lancelot, 1978). The latter viewpoint was based on the orthodox interpretation of the Rb–Sr dating of the underlying Precambrian II-2 deposits, but this dating appears to indicate the age of superimposed metamorphism. The stromatolite composition fits the Middle Riphean age of the Precambrian II-3, though it does not exclude the Early Karatavian age of the unit. We arbitrarily assign these stromatolites to the undivided Middle Riphean. But in any

case, the Middle Riphean in Africa demonstrates extremely low, if not zeroth, stromatolite diversity.

As noted, the Upper Riphean stromatolite-bearing deposits are widespread in Africa. First of all, these are the Atar and Tifounke groups in the northern margin of the Taoudeni syneclise and Mauritanian Adrar; their counterparts in West Ahaggar (the Stromatolite Group) and Morocco; carbonate formations of the Katanga (the Roan and Lower Kundelungu groups), West Congo and Lindian supergroups; the Bushimay, Bukoban, Musindozi and other groups of Equatorial Africa; and the Otavi and Gariep groups in the southwestern part of the continent (Bozhko, 1984; Bertrand-Sarfati, 1972a, 1972b, 1972c, 1976; Goodwin, 1991; Trompette, 1994).

The stromatolites of the Atar, Tifounke and Stromatolite groups are the best studied. According to Bertrand-Sarfati, they include several successive taxa sets which, according to their composition, can be grouped into two assemblages. *Conophyton ressoti* Mensh., *Con. jacqueti* Bertr.-Sarf., *Jacutophyton sakhariensis* Bertr.-Sarf., *Parmites concrescens* Raab., *Baicalia safia* Bertr.-Sarf., *B. lacera* Semikh. (=*B. maurltanica* Bertr.-Sarf), *Tungussia glebulosa* Bertr.-Sarf., *Gymnosolen hankii* Bertr.-Sarf., and some other forms are typical of the lower assemblage from units I_5–I_6 of the Atar Group and their correlatives. *Jurusania lissa* Bertr.-Sarf., *J. nizvensis* Raab., *Gymnosolen directus* Bertr.-Sarf., *Tarioufetia hemisphaerica* Bertr.-Sarf., *Tungussia nodosa* Semikh., *Tifounkeia ramificata* Bertr.-Sarf. are characteristic of the upper assemblage occurring in units I_7–I_{11} of the Atar Group and in unit I_{12} of the Tifounke Group (Bertrand-Sarfati, 1972a; Bertrand-Sarfati and Moussine-Poushkine, 1985). It should be stressed that both assemblages include genera and some species typical of the Upper Riphean of Northern Eurasia.

A number of Rb–Sr isochron datings on fine illite fractions from the Atar Group and from the unconformably preceding Shar Group, are as follows: 998 ± 34 Ma for the Shar Group; 890 ± 37, 874 ± 23, 866 ± 70, and 775 ± 54 Ma for units I_5, I_6, I_7, and I_{10} of the Atar Group, respectively (Clauer *et al.*, 1982). The Stromatolite Group of West Ahaggar includes correlatives of the lower part of the Mauritanian Adrar section—up to unit I_6, or to the lower beds of the unit I_7 (Bertrand-Sarfati, 1972a; Trompette, 1994). The time range of this group is determined by the Rb–Sr ages of unconformably underlying rhyolites (1075 Ma) and cutting intrusions (785 ± 55 Ma) (Cahen *et al.*, 1984). The taxonomic composition of the listed assemblages and isotopic datings (with due regard for analytical errors) suggest that the first assemblage is related to an upper part of the Lower Karatavian, and the second one corresponds to the entire Upper Karatavian.

Stromatolites of the lower part of the Lower Karatavian are known from the Roan Group, and comparable Bushimay, Musindozi, Lindian, Bukoban and other groups of Equatorial Africa, though they have been inadequately studied (Raaben and Zabrodin, 1972; Bozhko *et al.*, 1974; Bertrand-Sarfati, 1972b, 1976; Cahen, 1972). Several characteristic representatives of the first assemblage mentioned above, together with some local species of interprovincial genera, are reported from these units. Their fossiliferous deposits rest unconformably on the 1100-Ma-old rocks, and are overlain by tillites of the Great Conglomerate deposited ca. 950 Ma ago (Cahen *et al.*, 1984; Trompette, 1994). The taxonomic composition of stromatolites from the Roan Group evidences its Late Riphean age, while the available age measurements demonstrate that they belong to the very beginning of the Late Riphean. Stromatolites from the Roan Group include characteristic forms of the first assemblage typical of the Atar Group, thus implying that this assemblage characterizes the entire Early Karatavian.

Outside the Taoudeni syneclise and West Ahaggar, the Late Karatavian stromatolites of Africa are only known from the Schisto-Calcaire Group of the West Congo Supergroup, and the Otavi Group of the southwestern part of the continent. *Linella minuta* Bertr.-Sarf., *Jurusania* sp., *Tungussia congoliana* Bertr.-Sarf., and *Columnaefacta monbeliana* Bertr.-Sarf. were described from the first group (Bertrand-Sarfati, 1972a, 1972c; Bertrand-Sarfati and Vicat, 1987), and *Baicalia* aff. *rara* Semikh. and *Conophyton ressoti* Mensh. from the second one (from the Abenab Subgroup; Cloud and Semikhatov, 1969). Obviously, the lists do not reflect the actual diversity of all the forms. According to the isotopic-geochronological and chemostratigraphic evidence (Kaufman *et al.*, 1991), the Otavi Group belongs to an upper part of the Upper Karatavian (younger than 728 ± 40 Ma), whereas the Schisto-Calcaire Group corresponds to its lower part (older than 750–770 Ma; Cahen *et al.*, 1984).

The above data show that the Early Karatavian in Africa demonstrates a noticeable increase in diversity and a lateral expansion of stromatolites, while the Late Karatavian was an interval with a minor decrease in stromatolite diversity against the background of significantly reduced lateral distribution (Fig. 1). The diversity decrease would be more evident, assuming that stromatolites from the unit I_7 having the Rb–Sr age of 866 ± 70 Ma belong to the Early Karatavian (the age of the upper boundary of this unit is 850 Ma). However, the age of the unit I_7 was determined with a considerable error, and the stromatolite composition is more consistent with its Late Karatavian age.

The Vendian stromatolites in Africa, which do not have a distribution as broad as the Late Riphean forms, are recorded from three regions only: the Nama River basin (Gresse and Germs, 1993), the Volta depression (Bozhko *et al.*, 1974), and the Anti-Atlas (Semikhatov, 1977; Schmitt, 1978, 1979a, 1979b; Raaben, 1980). There are no data on the stromatolite taxonomic composition of the first of the regions; only one taxon was identified in the second region, and in the third region

the position of the Vendian–Cambrian boundary remains debatable (for bibliography see Bertrand-Sarfati, 1981; Latham and Riding, 1990). As it is accepted now, the lower part of the Adoudounian (with basal formation and Lower Dolomites overlying magmatic rocks having the U–Pb zircon age of 578 ± 15 and 563 ± 20 Ma; Cahen *et al.*, 1984; Compston *et al.*, 1992) belongs to the Vendian. As for the upper part of the Adoudounian (the Lie-de-Vin "Series"), its age is treated differently by different authors. Recent chemostratigraphic, paleontologic and partly paleomagnetic data show that the Cambrian lower boundary in the Anti-Atlas section should be placed near the top of the Lower Dolomites, and therefore, the Lie-de-Vin "Series" is Tommotian in age (Latham and Riding, 1990; Magaritz *et al.*, 1991; Compston *et al.*, 1992), as was suggested earlier by some stromatolite researchers (Semikhatov, 1977; Bertrand-Sarfati, 1981). If so, *Paniscollenia* sp. is the only taxon present in the Vendian of Morocco, while the Lower Cambrian deposits contain a rich stromatolite assemblage including, in addition to endemics, *Parmites victorius* Sid. (=*Linella* (?) *avis*, sensu Schmitt, 1979a), *Tungussia* cf. *inna* Walt., *Patomia* sp., local species of *Acaciella*, *Vetella* and *Tifounkeia*, as well as *Madiganites* in the High Atlas. The first three taxa are the only forms confined to the controversial (Vendian or Cambrian) interval of the section. Thus, according to present knowledge of the African stromatolite taxonomy, the Vendian is marked by a considerable decrease, and the Lower Cambrian, by an increase in diversity (Fig. 1). But these trends should be refined through the studies of rather abundant (Gresse and Germs, 1993) Vendian stromatolites of the Nama basin.

Thus, the histograms for African regions (Fig. 1) exhibit generally low variations in the Proterozoic stromatolite diversity and, unlike other superregional histograms, a zero depression corresponding to the entire Early Proterozoic. The absence (in any case, scarcity) of stromatolites in the Lower Proterozoic deposits of the continent was caused by their specific composition and sedimentation environments. The lower part of the sedimentary successions accumulated here prior to the Eburnian tectonic-thermal event (ca. 2.1–1.95 Ga); it is dominated by siliciclastic and volcanogenic rocks deposited in isolated basins or in relatively deep-water settings of the extensive mobile belts. If present, the carbonate units of this part of the sections (e.g., of the Francevillian Group of Equatorial Africa, 2.14–2.05 Ga; Trompette, 1994) lack stromatolites. The Eburnian event resulted in the consolidation of the continental craton and its general uplift. Against the background of these events, the molasses and subaerial volcanics accumulated in some isolated depressions (Khain and Bozhko, 1988; Cahen *et al.*, 1984; Trompette, 1994).

During the Early Riphean, most of the continent represented an elevated area, but some intracratonic mobile belts originating at that time accumulated some carbonate units, including microbiolite-bearing ones. Stromatolites from the most significant of the above units (the upper group of the Kibaran Supergroup) have yet not been studied. Therefore, the slight stromatolite diversity increase, which is exemplified by taxa from the Precambrian II-2 of Morocco, reflects the tendency in the compositional changes but not their actual amplitude. The Middle Riphean depression in the histogram under consideration mirrors the restricted distribution and predominantly clastic, or volcaniclastic composition of the relevant deposits in Africa. This phenomenon is usually considered to be related to the termination of the Kibaran orogeny at the beginning of the Middle Riphean (ca. 1.3 Ga), the wide distribution of orogenic molasses, and the subsequent general tendency to the continental uplift (Khain and Bozhko, 1988; Cahen *et al.*, 1984; Goodwin, 1991; Trompette, 1994).

The Early Karatavian peak in the histogram of the African stromatolite diversity is correlated with two events which affected the entire continent: (1) the wide marine transgression over the stable massifs, which flooded almost the entire western part of the continent and some extensive areas in its equatorial part (the Congo syneclise); (2) the widespread carbonate sedimentation in the intracratonic mobile belts (those of Katanga, Pharusian, West Congo, Damara and others) formed at the very end of the Middle Riphean or at the very beginning of the Late Riphean in Equatorial and southwestern Africa (Khain and Bozhko, 1988; Goodwin, 1991; Trompette, 1994). It is noteworthy that the Lower Karatavian carbonate deposits are mainly represented by limestones, whereas those of the over and underlying horizons of the African Precambrian are dominated by dolomites.

Of great interest are the chronological relationships between the African stromatolite assemblages and two horizons of the Upper Riphean tillites (the Great and Small Conglomerates of the Katanga Supergroup and their counterparts) deposited during the first phases of the African Glacioera (Chumakov, 1978). Regretfully, in the area of the tillite distribution, the Late Riphean stromatolites are abundant and to some extent studied only in the deposits underlying the Great Conglomerate (the Roan, Bushimay and other groups); they are rare and almost unstudied in the deposits between or above the tillite horizons. Therefore, an understanding of the mentioned relationships requires the correlation between tillite-bearing sections of Equatorial Africa and the predominantly carbonate successions of West Africa containing abundant and well studied stromatolites. These correlations, based on isotopic-geochronologic and partly regional geological data (Trompette, 1973, 1994; Cahen *et al.*, 1984), show that the tectonic movements prior to the lower tillites and this horizon proper (ca. 950 Ma) correspond in time to the hiatus between the Shar and Atar groups and to the basal coarse clastic sediments of the latter group; the upper tillite beds (younger than 870–850 Ma and older than 814 ± 15 Ma) are likely correlative with basal horizons

of the deposits containing the Late Karatavian stromatolite assemblage. If this is the case: (1) the lower tillites are inside the stratigraphic interval of the Early Karatavian stromatolite assemblage; (2) a wide expansion of this assemblage into western Africa proceeded under the conditions of a post-glacial transgression; (3) the Late Karatavian assemblage of the African stromatolites formed against the background of the Riphean glacier retreat from this continent. It should be noted that units I_6 and I_7 of the Atar Group yield stromatolites with a specific type of microstructure, whose origination is considered to be an important biotic event related to changes in sea water composition (Fairchild *et al.*, 1990; Bertrand-Sarfati and Pentecost, 1992).

The slight decrease in the African stromatolite diversity during the Late Karatavian was against the background of a distinct reduction of depositional areas, except for some basins, and the general reduction of carbonate sedimentation then replaced by siliciclastic sedimentation prevalent on continents since the very end of the Late Riphean, 770–750 Ma ago. These events are considered as the results of the Katangan (Lusakian) orogeny (Khain and Bozhko, 1988; Cahen *et al.*, 1984; Goodwin, 1991).

The sharp Vendian decrease in African stromatolite diversity (Fig. 1) seems to reflect, at least to some extent, the poor state of relevant studies. Nevertheless, the fact is that a noticeable contraction of sedimentation areas during the Vendian over the whole continent was a response to the Katangian and Pan-African events, and few still active depressions, except for the Nama basin, are poor in carbonates, especially in their stromatolitic varieties (Gresse and Germs, 1993; Trompette, 1994).

Thus, the general trend of changeable diversity of the African stromatolites, which is depicted from available but incomplete records, appears to be specific in the Lower Proterozoic part, while its Riphean–Vendian part reveals a kind of similarity to the pattern based on the better studied stromatolites of Northern Eurasia and China. This resemblance is evident from the higher diversity in the Early Burzyanian, and especially in the Early Karatavian time and from the diversity reduction at the end of the Proterozoic. At the same time, histograms for Africa and India display the Late Burzyanian–Yurmatinian minimum, which reflects the scarcity of stromatolites of that age in both superregions.

Australia

In the relatively small territory of Australia, Proterozoic stromatolites occur in a number of lithostratigraphic units scattered throughout the section (Grey, 1982; Plumb, 1985, 1990). Their best known localities are in the following structures of different age: the Early Proterozoic Hamersley, Capricorn, Glengarry, and Earaheedy basins in the west of the continent; the Early Riphean McArthur basin and the Middle Riphean Bangemall depression in the north and west, respectively; and the Late Riphean–Vendian Amadeus basin and the Adelaide "geosyncline" in the central and southern parts of Australia. The valid description and identifications of these microbiolites have appeared in the literature over the last 25 years (Glaessner *et al.*, 1969; Cloud and Semikhatov, 1969; Walter, 1972; Preiss, 1972, 1973, 1974, 1976; Walter *et al.*, 1979, 1988; Grey, 1978, 1981, 1984, 1985a, 1985b, 1994a, 1994b; Preiss and Krylov, 1980; Grey and Thorn, 1985; Gee and Grey, 1993). Earlier publications (e.g., Edgell, 1964) are of historical interest only. The main contribution to studies on the Australian stromatolites was made by researchers who followed uniform approaches to classification.

The Proterozoic stromatolites from Australia are much better studied than those from Africa and India. The sample discussed below consists of 75 taxa of a specific level representing 42 formal genera (groups). Without question, these figures do not represent the actual diversity, because the stromatolites from some Australian areas still remain undescribed, and the studied collection is dominated by columnar forms and partly by microstromatites. Nevertheless, we can state with reasonable confidence that the Precambrian stromatolites of Australia are generally less diverse than those of Northern Eurasia and China.

The oldest stromatolites of the continent are known from the Lower (the Warrawoona Group, 3.56 Ga) and Upper Archean, but they occur sporadically (Grey, 1981; Walter, 1983). Among their columnar-stratiform, nodular and stratiform buildups, only three species of three genera are supplemented with valid descriptions (Cloud and Semikhatov, 1969; Walter, 1972; Grey, 1981). Stromatolites from the Warrawoona Group of Australia and the Onverwacht Group of Africa (3.54 Ga) used to be considered as the oldest microbiolites, but recently their biogenic origin has been in doubt (Lowe, 1994).

The oldest Proterozoic stromatolites of Australia are recorded from the Hamersly and Ture Creek groups from the northwestern part of the continent (Edgell, 1964; Walter, 1972; Grey, 1982; Plumb, 1985), but they lack a valid description. According to Grey (personal communication), the present columnar branching buildups, which should be considered as a new genus and species associated with specific microstromatites; it may be arbitrarily suggested that these buildups represent two species and two genera. The respective host deposits represent the first unit of the adopted Proterozoic scale: they include the volcanics with the U–Pb zircon age of 2490 ± 20–2434 ± 18 Ma and are underlain by the Fortesque Group, which is 2765–2687 Ma old (Arndt *et al.*, 1991; Pidgeon and Horwitz, 1991).

The deposits 2.3–2.0 Ga in age have not been distinguished in Australia (Plumb, 1990). The younger Early Proterozoic deposits of several areas in the western and northwestern parts of the continent contain stromato-

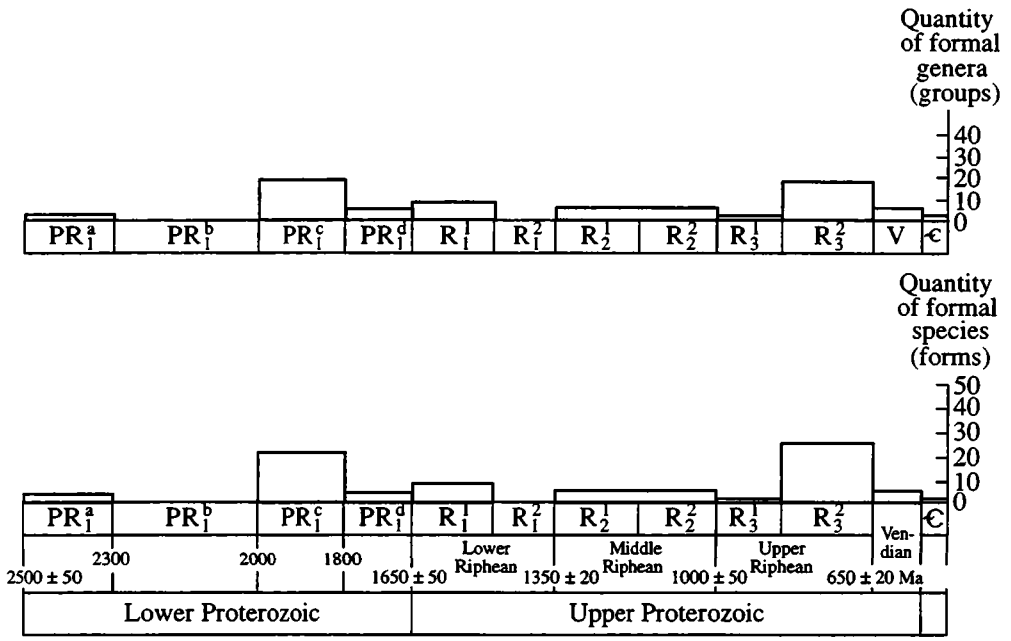

Fig. 2. Secular variations in diversity of Proterozoic stromatolites from Australia at the levels of formal species (forms) and formal genera (groups).

lites. Their taxonomic composition is studied only in the Wyloo (the Capricorn orogen), Glengarry and Earaheedy (the Glengarry and Earaheedy basins considered earlier as parts of the single Nabberu basin), South Alligator (the Pine Creek "geosyncline"), Kimberley (the Kimberley basin), Capricorn (the Blair basin), and Haslingen groups (Mt. Isa Province). Not all the groups have strict age constraints.

The South Alligator and Wyloo groups are definitely referred to the third Proterozoic subdivision (2.0–1.8 Ga = Orosirian). The stromatolite-bearing Copley Formation of the South Alligator Group is conformably overlain by tuffs with a U–Pb zircon age of 1884 ± 3 and 1877 ± 11 Ma (Needham *et al.*, 1988). The Wyloo Group is intruded by granodiorites with the Rb–Sr age of 1680 ± 30 Ma (mineral isochron), and its main stromatolitic horizon, the Duck Creek Dolomites, conformably underlies volcanics with a U–Pb zircon age of 1843 ± 2 Ma (Thorne and Seymour, 1991; Pidgeon and Horwitz, 1991).

The time range of the Glengarry Group is also estimated to be 2.0–1.8 Ga (Plumb, 1990; Gee, 1990; Thorne and Seymour, 1991; Gee and Grey, 1993). This group, younger than the Upper Archean granites (about 2.65 Ga), includes galenite mineralization with the model Pb–Pb age of 1.8 Ga, and is correlative to the Wyloo Group. The age of the Earaheedy Group resting unconformably on the Grengarry Group is less definitely constrained. The underlying granitoids have an age of 2400 Ma (U–Pb), while intrusive syenites are 1630 Ma old (Rb–Sr method; Plumb, 1985, 1990). Apparently, the Rb–Sr and K–Ar ages of 1700–1670 Ma, which were measured on glauconite and illite from this group, define its minimal age limit (Bunting, 1986). According to these data, the age of the Earaheedy Group is usually suggested to be 1.7–1.65 Ga (Preiss, 1976; Plumb, 1985, 1990; Gee, 1990). However, paleontological remains seem to be able to refine this estimation. The silicified microfossils from this group include typical representatives of the Gunflint assemblage (for bibliography see Grey, 1994a), whose age is approximated at 2.0 Ga (Moore *et al.*, 1992), and among stromatolites of the group, as Grey notes, there are forms similar to those confined to the deposits older than 1.8 Ga. These facts allow us, as does Grey (1994a), to suggest that the Earaheedy Group was accumulated ca. 1.8–1.9 Ga ago.

Stromatolites of the Wyloo Group are dispersed through several formations, but they are abundant and well described in the Duck Creek Dolomite only (Cloud and Semikhatov, 1969; Walter, 1972; Preiss, 1976; Grey, 1985b). The unit includes columnar *Pilbaria perplexa* Walt., and *Conophyton* sp., along with microstromatites *Asperia ashburtonia* Grey, *Yelma digitata* Grey, and some others. *Pilbaria perplexa* Walt., local representatives of *Kussoidella* and *Segosia, Conophyton* sp. indet., microstromatites *Katernia perlina* Bertr.-Sarf. et Erick. and some others, including *Thyssagetes*-like buildups (Gee and Grey, 1993; Grey, 1994b), are known from the South Alligator and Glengarry groups. The Earaheedy Group contains more diverse stromatolites (Gee and Grey, 1993; Grey, 1994a) represented by *Yelma digitata* Grey, local species of *Pilbaria, Eocapsiphora, Externia, Omachtenia, Ephyaltes* and several other interprovincial and endemic genera. Thus, the deposits of 2.0–1.8 Ga age

in Australia yield a diverse stromatolite assemblage (Fig. 2), whose typical forms are *Pilbaria perplexa* Walt., and other species of this genus associated with particular microstromatites.

Australian stromatolites assigned to the terminal unit of the Early Proterozoic (1.8–1.65 Ga) were described from the Haslingden, Kimberley and Capricorn groups. Stromatolite-bearing dolomites of the first group are cut by granites with an age of 1740–1730 Ma and conformably overlie the 1790 ± 9-Ma-old volcanics (U–Pb, zircon; Page, 1988). The Kimberley Group unconformably overlies the volcanics with zircon age of 1850 ± 5 Ma, while the dolerites intruding these deposits have the Rb–Sr age of 1760 ± 25 Ma (Page, 1988). The Capricorn Group is cut by the aforementioned granodiorites with the Rb–Sr age of 1680 ± 30 Ma, and unconformably overlies the Wyloo Group (1843 ± 2 Ma and older, see above). Stromatolites are rare in these groups: *Kussoidella* sp., *Conophyton* sp., and *?Segosia isa* Walt. et Kryl. are the only ones identified among them (Grey, 1982; Walter *et al.*, 1988).

Thus, the histogram of the Australian stromatolite diversity (Fig. 2) shows a zero depression over the interval of 2.3–2.0 Ga, which is followed by a noticeable peak defined by the taxa from the Glengarry, Wyloo, South Alligator, and Earaheedy groups (2.0–1.8 Ga), and then a diversity decrease corresponding to the stromatolite composition in the Haslingden, Kimberley and Capricorn groups (1.8–1.65 Ga). If the age of the Earaheedy Group is not 1.8–1.9 Ga, as we accepted, but ca. 1.7 Ga, as it is assumed by many Australian colleagues, the diversity peak of the Early Proterozoic stromatolites of Australia will move from the third to the fourth (terminal) unit of the Early Proterozoic because of the age constraints for a single particular basin.

The stromatolite-bearing deposits corresponding in age to the early Early Riphean are rather widespread in Australia and are concentrated mostly in the North Australian craton (the McArthur and Nathan groups of the McArthur basin, the Birrindudu Group and Bungle-Bungle Dolomite of the Birrindudu basin, the McNamara Group of the Mt. Isa Province, the Scorpion Group and others). Only a few stromatolites from the McArthur, Nathan and McNamara groups have a valid description (Cloud and Semikhatov, 1969; Walter, 1972; Walter *et al.*, 1988). These are *Kussiella kussiensis* Kryl., *Omachtenia omachtensis* Nuzhn., *Conophyton garganicus* Kor. (Walter *et al.*, 1988) and *Kussiella vitata* Kom. (identified by Semikhatov from the Walters collection), and are all characteristic of the Early Riphean of Eurasia. Besides, there are Thyssagetacean buildups of *Australoconus abnera* Walt. et Kryl., the earliest representatives of the genus *Acaciella*, and some others (specimens of *Eocapsophora paradisa* Cloud et Semikh.) have also been reported from the McNamara Group.

The Early Burzyanian age of the McArthur and Nathan groups is defined by the following: (1) the Nathan Group is unconformably overlain by the Roper Group with a glauconite age of 1250–1400 Ma (K–Ar, model Rb–Sr) and illite age of 1429 ± 31 Ma (Rb–Sr isochronous; Kralik, 1982); (2) the U–Pb zircon age of tuffs from the basal horizons of the McArthur Group below its stromatolite-bearing carbonate succession and from the correlatives of the group in the Mt. Isa Province is equal to 1690 ± 27 and 1670 ± 19 Ma; (3) the U–Pb and Rb–Sr ages of granites underlying this group are 1610 ± 10–1698 ± 25 and 1688 ± 13 Ma, respectively (Page, 1988). The McNamara Group was separated by a metamorphic event from the granites with the U–Pb age of 1740 ± 24 Ma (Plumb, 1990), and is correlated with a lower part of the McArthur Group. The other listed stromatolite-bearing groups are correlated by geological evidences with the McArthur and/or Nathan groups (Plumb *et al.*, 1980; Plumb, 1990), but there are no data on their stromatolite composition, except for the finds of *Conophyton* sp. from the Scorpion Group (Grey, 1982).

Unmetamorphozed successions of the Upper Burzyanian and Yurmatinian (the upper Lower and Middle Riphean) have a limited distribution in Australia. They are usually composed of clastic and/or volcanogenic rocks and contain few, if any, stromatolite-bearing carbonates (Walter, 1972; Grey, 1982; Plumb, 1985). Valid identifications of these stromatolites are available for the Tolmer and Bangemall groups and their correlatives in the northern and northwestern areas of the continent. The Tolmer Group contains a form originally described as *Inzeria tjomusi* Kryl. (Cloud and Semikhatov, 1969), but this identification was later doubted (Walter, 1972; Grey, 1982). The Bangemall Group and its correlatives yield abundant but taxonomically monotonous stromatolites represented by *Baicalia capricorna* Walt., *Conophyton australe* Walt., *Colonnella* sp., *Acaciella* cf. *australica* Walt., and others (Walter, 1972; Grey, 1982, 1985a). According to Grey's opinion, the latter form should be identified as a new species.

For glauconites from the Tolmer Group, the isotopic K–Ar and Rb–Sr ages approximate to 1200 Ma (Webb and Page, 1977) and seem to be its only numerical age determinations. The Rb–Sr isochron ages on the whole-rock samples of volcanics and shales from the Bangemall Group are equal to 1098 ± 42 and 1080 ± 80 Ma, respectively, and it was accepted that the group is ca. 1100 Ma old (Compston and Arriens, 1968; Grey, 1985a; Plumb, 1990). However, the model Pb age of the apparently syngenetic galenite from the lower part of the group appeared to be about 1500 Ma (Chuck, 1984), and the rhyolites near its base are even older (U–Pb data for zircon, SHRIMP; Knoll personal communication), partly below the maximal age limit of the group defined by its relations with the underlying rocks. As for the Bangemall Group stromatolites, their composition suggests the Middle or Late Riphean age.

Taken as a whole, these data allow us to assign the Tolmer and Bangemall groups to the Middle Riphean in general.

The problem of the diversity dynamics of the Australian stromatolites at the end of the Riphean is directly dependent on the interpretation of the time range of the Adelaide Supergroup of South Australia and correlative deposits in the central part of the continent (in the Amadeus, Officer, Georgina and Ngalia basins), which contain a considerable amount of valid taxa. The upper part of this supergroup, containing the famous Ediacarian fossils and unconformably underlying the Lower Cambrian deposits, is definitely of the Vendian age, whereas its lower part was long considered, on the basis of some miscorrelations and indirect assumptions, as embracing the stratigraphic interval down to 1100, and even 1400 Ma (for bibliography see Keller *et al.*, 1984; Preiss, 1987). These conclusions were revised in accordance with new evidence.

The concordant U–Pb zircon age for the Rook tuffs of the Curdimurka Subgroup (the second subgroup of the Callanna Group, the oldest in the Adelaide succession) is the most important age constraint for the lower part of the supergroup. This age, equal to 802 ± 10 Ma (Fanning *et al.*, 1986), proves that the Kurdimurka Subgroup and respective part of the Adelaide Supergroup above are younger than 800 Ma (Preiss, 1987; Plumb, 1990). The lower part of the Callanna Group below the Kurdimurka Subgroup is unlikely to be much older, because there is no sign of considerable hiatuses between both units. This is confirmed by the Sm–Nd mineral isochron age (802 ± 35 and 865 ± 49 Ma) for basic dykes preceding the rifting stage in the Adelaide "geosyncline" (Drexel *et al.*, 1993). Thus, the Adelaide Supergroup is younger than 850 Ma (Plumb, 1990; Drexel *et al.*, 1993), and belongs to the Late Karatavian and Vendian.

The Upper Precambrian deposits of the Amadeus basin and adjacent depressions of Central Australia are also referred to the Upper Karatavian and Vendian on the basis of their general similarity and use, for the upper part of the section, direct lithostratigraphic correlation with the Adelaide Supergroup (Preiss, 1987). The Rb–Sr isochron age for the Stuart Dyke Swarm (897 ± 9 Ma; Black *et al.*, 1980) antedating the formation of the Amadeus basin once was used to support such a conclusion. However, the Sm–Nd isochron age of that dyke swarm appeared to be 1061 ± 36 Ma (Zhao *et al.*, 1992). Now the Callanna and/or Burra groups constituting the lower part of the Adelaide Supergroup are taken in Australian literature as an analogue of the Heavitree and Bitter Springs formations, the lowermost units in the Riphean succession of the Amadeus basin.

Thus, the Upper Precambrian deposits of southern and central Australia (the Adelaide Supergroup and its correlatives) containing abundant and relatively well studied stromatolites belong to the Upper Karatavian and Vendian. The lower Vendian boundary can be placed at the base of the tillites of the Yerelina Subgroup, the upper subgroup of the Umberetana Group, and coeval tillites of the Amadeus basin. This placing of the boundary is substantiated by the stratigraphic position of the Ediacarian fauna, Rb–Sr datings, and correlation of the tillite horizons (Keller *et al.*, 1984). As for the Lower Karatavian deposits, they have a limited distribution in Australia and, as far as we know, lack any identified stromatolites. The only possible exception is the so-called Robert succession of West Australia, which contains *Baicalia burra* Preiss, and is provisionally regarded as 1.0–0.9 Ga old (Grey, personal communication).

The Upper Karatavian deposits, which form a part of the Adelaide Supergroup and its correlatives in the Amadeus basin, yielded more than one third of the species and nearly a half of the stromatolite genera known from all of the Proterozoic deposits of Australia (Glaessner *et al.*, 1969; Walter, 1972; Preiss, 1972, 1973, 1974, 1987; Preiss and Krylov, 1980). Among them, there are some typical Late Riphean genera (*Boxonia, Gymnosolen, Inzeria, Jurusania, Katavia, Linella*, and *Minjaria*), which are represented by endemic and interprovincial species (*Inzeria tjomusi* Kryl., *Jurusania nizvensis* Raab., and *Linella avis* Kryl.). They associate with *Conophyton garganicus* Kor., endemic representatives of *Baicalia, Tungussia* and other interprovincial and endemic genera. Though rich, this list does not exhibit the actual diversity of the Late Karatavian stromatolites in Australia, because the stratiform and nodular buildups from the Adelaide Supergroup and correlative deposits of central Australia, as well as all stromatolite morphotypes of the coeval Yanina Group of West Australia, have yet to be described in literature.

The Vendian deposits of Australia are poor in stromatolites. The latter are represented by endemic species of *Linella, Tungussia, Elleria*, and *Jacutophyton* (=*Georginia*) described from the Georgina and Amadeus basins and from the Adelaide "geosyncline" (Walter, 1972; Grey, 1982; Preiss, 1987), and apparently by *Conophyton basalticum*, Walt. and *Con.* cf. *gaubitza* Kryl. from the Antrim Plateau basalts. These basalts are conventionally assigned in Australian literature to the Cambrian, but the occurrence of *Conophyton* is more consistent with their Vendian age. According to published data, the doubtless Cambrian deposits of Australia yielded only two stromatolite taxa.

The histogram of the Australian stromatolite diversity (Fig. 2) is similar to the previously described graphs for Indian and African superregions, as it has a generally flat pattern. Nevertheless, it shows obvious specific features: (1) a deep (down to zero over the interval of 2.3–2.0 Ga ago) depression corresponding to the first half of the Early Proterozoic; (2) the most distinct peaks of relative diversity displaced toward the third unit of the Early Proterozoic (2.0–1.8 Ga) and the Late Karatavian; (3) a very low taxonomic diversity

characteristic of the main interval of the Riphean from the Late Burzyanian to the Early Karatavian inclusive. The Australian histogram also demonstrates some features typical of histograms for other superregions, namely the diversity decrease at the very end of the Early and Late Proterozoic and the particular peak over the Early Burzyanian interval. Certainly, due to the general state of our knowledge on the Australian stromatolites, this histogram does not adequately reflect the real secular variations in abundance of their taxa; however, it seems to display the actual trends of increase and decrease in the taxonomic diversity of these stromatolites.

Going to the geological events associated with the discussed changes in stromatolite diversity, we have to mention that only a minor part of West Australia was stable during the Early Proterozoic, while the main part of the present-day continent underwent cratonization 1.88–1.84 Ga ago (the Barramundian orogeny) and later (Khain and Bozhko, 1988; Plumb *et al.*, 1980; Plumb, 1985; Page, 1988; Goodwin, 1991). The extremely low abundance of stromatolites in the first half of the Early Proterozoic of Australia (2.5–2.0 Ga) is a result of the sedimentation environments of that time, when minor epi-Archean stable massifs of West Australia tended to uprise, and mobile belts in the rest of the continent accumulated mostly volcanogenic and siliciclastic deposits. Only the Hamersley basin in West Australia, which was inherited from the Late Archean, showed a noticeable distribution of stromatolites at the very beginning of the Early Proterozoic.

A splash of diversity of the Australian stromatolites 2.0–1.8 Ga ago was a direct result of the short-range development of riftogenic depressions, which were of the intracratonic type in West Australia (the Wyloo, Glengarry, and, likely, Earaheedy basins), and contained abundant stromatolites. The concurrent mobile belts of central and northern Australia were poor in carbonate deposits, especially in those bearing stromatolites. The decrease in stromatolite diversity at the very end of the Early Proterozoic coincided with two successive events which were of continental scale: (1) the Barramundian orogeny (1880–1840 Ma ago), which resulted in the formation of a characteristic acid volcano-plutonic association and in cratonization and rising of the North Australian platform, and (2) the short-term development (from 1800–1790 to 1750–1740 Ma) of a new series of riftogenic depressions over the epi-Barramundian basement (the Pine Creek, Kimberley and others), which were filled by volcaniclastic deposits with rare carbonate members only (Page, 1988; Wyborn *et al.*, 1988; Goodwin, 1991).

The wide Early Burzyanian marine transgression spread predominantly over the North Australian craton and left behind shallow-water siliciclastic-carbonate deposits of the McArthur and Nathan groups and their correlatives accumulated under the warm and arid climatic conditions. The diversity of stromatolites described from these deposits (Walter, 1972; Walter *et al.*, 1988) is clearly reflected in the histogram (Fig. 2). As for the extensive deep depression (with a minor peak in the middle) corresponding to the main part of the Riphean, it reflects a general tendency of the continental uplift during the Late Burzyanian–Early Karatavian time. Sedimentation of that time was concentrated in some mobile belts or isolated intracratonic basins, and had a predominantly siliciclastic or volcanogenic-siliciclastic character. The specific features of one of these basins (Bangemall) were responsible for the appearance of abundant, though low-diverse stromatolites there. The so-called Musgrave event at the very end of the Middle Riphean resulted in the stabilization of the Central Australia mobile belts and general emergence of the area with the related areal reduction of the Early Karatavian sedimentation (Plumb, 1990; Goodwin, 1991).

Principal Late Karatavian changes in the paleogeography of the continent, such as development of the Adelaide, Amadeus and other riftogenic structures and following marine transgression over the adjacent shelves (Preiss, 1987; Plumb, 1985, 1990; Drexel *et al.*, 1993), as one of the consequences, caused the deposition of relatively abundant stromatolite-bearing carbonates. The taxonomically diverse stromatolite buildups were formed in a wide range of settings (Preiss, 1987), and this resulted in the highest diversity peak of the Australian Proterozoic stromatolites. The Sturtian tillites, widespread in southern and in parts of central Australia, happened to be deposited within the time range of this peak. Below and above these tillites, the Late Karatavian stromatolites of the Adelaide "geosyncline" are different in composition, as they are more diverse in the post-tillite part of the succession.

The Vendian diversity decrease of Australian stromatolites was against the background of such events as the two-phase postglacial transgression, predominance of the siliciclastic deposition, and the basic volcanism at the very end of the Vendian (the Antrim Plateau basalts). Noticeable amounts of Vendian carbonates are known only in the Adelaide Supergroup, where they compose the cap-dolomites of the Nuccaleena Formation and constitute Wonoka Formation underlying the Pound Quartzite. The first unit lacks stromatolite, and the second one shows very scarce stromatolite buildups in spite of the depositional environments favorable for their growth (Preiss, 1987). In this case, the reduction in stromatolite abundance was very likely due to the destructive activity of Metazoa, which used to be considered as the main cause of stromatolite decline at the end of the Proterozoic (Awramik, 1971, 1992; Walter and Heys, 1985). In any case, the diversity dynamics of the Proterozoic stromatolites of Australia was under the control of interacting different-scale factors. It is another matter that, due to the small size of the continent, effects of regional and superregional factors were not uncommon in the foreground. This seems to explain the fact, noted in literature (Grey, 1982), that the diversity dynamics of Australian stromatolites is different from that recognized on the basis of larger samples.

North America

The North American Precambrian stromatolites demonstrate a wide range of stratigraphic distribution from the Upper Archean to the Vendian inclusive, but generally they are of low abundance in the sections. They were objects of paleontological description since the end of the last century. In the early 20th century, they were intensively studied by the pioneers of Precambrian paleontology C. Walcott, C. Fenton, and M. Fenton, who were striving to reveal their taxonomic composition and stratigraphic significance. Later on, the American geologists became sceptical with regard to the classification and stratigraphic potential of stromatolites, and only occasionally did they treat them as paleontological objects (Rezak, 1957). The present situation has not changed significantly. Therefore, the understanding of the taxonomic composition of the American stromatolites remains unsatisfactory. The first investigators did not take into consideration some diagnostic features significant in the modern systematics, and up-to-date descriptions are scarce and dedicated to selected morphotypes from a few areas.

The presented diversity analysis of the American stromatolites is based on: (1) the published valid taxonomic descriptions from the standpoint of the modern classification and available adequate morphologic description and/or identifications substantiated by necessary investigations (Could and Semikhatov, 1969; Hofmann, 1969, 1974, 1978; Knight, 1968; Howell, 1971; Semikhatov, 1974, 1978; Bertrand-Sarfati and Caby, 1976; Horodyski, 1976, 1983; Aitken *et al.*, 1978a, 1978b; Weber *et al.*, 1979; Hofmann *et al.*, 1980; Cevallos-Ferriz and Weber, 1980; Cevallos-Ferritz *et al.*, 1988; Bertrand-Sarfati and Awramik, 1992), (2) Semikhatov's unpublished data derived from his study of the type collections of the Proterozoic stromatolites of the United States and identifications of some Canadian stromatolites. Stromatolite identifications known from literature and based only on field observations or comparison of selected buildup features (e.g., Donaldson, 1976; Hofmann, 1977, 1981) are treated as invalid and disregarded. In total, we considered the stratigraphic distribution of 74 taxa of a specific level representing 35 genera.

The oldest American stromatolites are reported from the Upper Archean of the Canadian shield and piedmonts of the Rocky Mountains in Wyoming (about a dozen localities; Hofmann *et al.*, 1991; Hofmann and Masson, 1994). They are represented by morphologically different morphotypes (including columnar and *Thyssagetes*-like forms and microstromatites) from commonly thin carbonate or chert units of the greenstone belt sections. The Steeprock Group of Canada and the Whalen Group of Wyoming only show Archean stromatolite-bearing carbonate units of a considerable thickness.

The first Proterozoic stromatolites of the continent are known from the lower part of the Huronian Supergroup of the Canadian shield. The thin carbonate Espanola Formation yields stratiform buildups of the *Stratifera* affinity and microstromatites with the same microstructure (Hofmann *et al.*, 1980). The time range of the Huronian Supergroup is detectable from the U–Pb age of the volcanics near its base (2450 ± 25 Ma), and that of the diabase cutting the entire supergroup (2219 ± 4 Ma; Krough *et al.*, 1984; Corfu and Andrews, 1986). Thus, these stromatolites can be assigned to the first unit of the adopted scale.

Amid the described North American stromatolites, the buildups of the Denault Formation of the Labrador trough (Donaldson, 1963; Semikhatov, 1978) and the Nash Fork Formation of the Medicine Bow Mountains, Wyoming (Knight, 1968) can be assigned to the second unit of this scale (2.3–2.0 Ga). The time range of the Denault Formation is defined by the U–Pb zircon ages of the deposits which conformably underlie (2142 ± 4 Ma) and unconformably rest upon it (1960–1880 Ma), and by the Pb–Pb age of galenite (1940–1980 Ma; Hoffman, 1988, 1989; Hoffman and Grotzinger, 1988). The Nash Fork Formation lies above the correlatives of the Gowganda tillites terminating the Huronian Supergroup (Young, 1973), and is considered to be older than the granitoids with the Rb–Sr age of 2000–2100 Ma (Houston *et al.*, 1992). *Asperia aspera* Semikh. and *Stratifera laxa* Semikh. were described from the Denault Formation, and some stratiform and small branching buildups from the Nash Fork Formation. Study of the Nash Fork stromatolites in the field and in Prof. Knight's collection has resulted in the identification of two taxa: *Stratifera* differing in microstructure from *Str. laxa*, and microstromatites similar to *Gruneria*. In addition, the large domes of *Hadrophycus immanus* Fent. et Fent. were described from these deposits (Fenton and Fenton, 1939). Thus, it can be conventionally assumed that the first unit of the Proterozoic in America contains two species and two genera of stromatolites, whereas the second unit yielded five species and four genera (Fig. 2).

The North American stromatolites are more widely distributed in deposits with an age range of 2.0–1.8 Ga, where they frequently constitute units of considerable thickness and extension. Stromatolites of that age are known from several areas of the Canadian shield (Douglas, 1972; Hofmann, 1981; Hoffman, 1988; Grotzinger, 1988a) but usually are not described in the literature in any details. Few publications present the general characteristics of their morphology (Donaldson, 1963; Cecile and Campbell, 1978), while valid descriptions are available only for a part of the forms present in the Epworth Group and the Great Slave Supergroup in the northwestern part of the shield, and for some buildups from the Mistassini, Animikie and Belcher groups in its southern and central parts (Hofmann, 1969, 1978; Cloud and Semikhatov, 1969; Semikhatov, 1978). Among the described forms, *Asperia aspera* Semikh., *Conophyton minusculum* Semikh., *Gruneria biwabikia* Cloud et Semikh., *Pilbaria perplexa* Walt., *Stratifera laxa* Semikh., *Mistassinia wabassion* Hofm. and *Kus-*

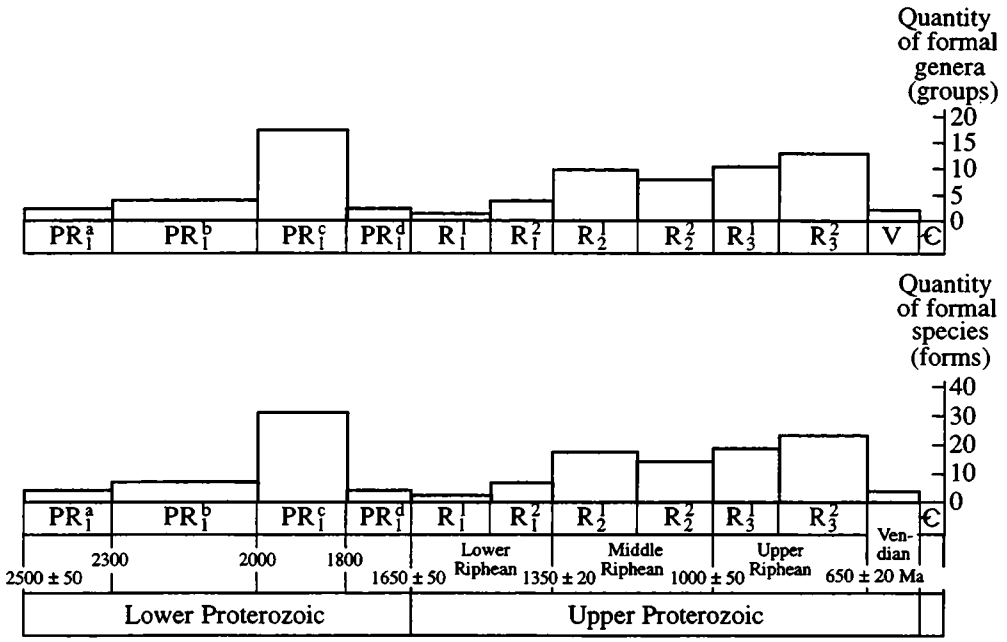

Fig. 3. Secular variations in diversity of Proterozoic stromatolites from North America at the levels of formal species (forms) and formal genera (groups).

soidella, which are reported from the Lower Proterozoic deposits of other superregions, attract attention. They are associated with a number of endemic taxa. In addition to the listed taxa, we should mentioned *Minjaria* sp. from the Rocknest Formation of the Epworth Group (identified by Semikhatov), and *Omachtenia* sp. and *Stratifera* sp. from the Gunflint Formation of the Animikie Group, the two latter forms with a specific microstructure (Awramik and Semikhatov, 1979). Thus, the diversity of the North American stromatolites increased noticeably 2.0–1.8 Ga ago in comparison with the previous time interval (Fig. 3).

The time ranges of the above units are discernible from the following. The stromatolite-bearing deposits of the Great Slave Supergroup lie between two tuff horizons with U–Pb zircon ages of 1969 ± 4 and 1882 ± 9 Ma, and are cut by granitoids 1860–1870 Ma in age (Hoffman, 1988). The Epworth Group is correlative with the Great Slave Supergroup (Hoffman, 1973), and its stromatolite-bearing units are younger than 1.92 and older than 1.90 Ga according to the U–Pb zircon datings of the tuffs (Grotzinger, 1988b). The basal part of the Animikie Group is lithostratigraphically correlated with volcanics with the U–Pb age of 1910 ± 10 Ma (the lower part of the Mennominee Group), and the whole group was affected by the Penokean orogeny with peak activity 1840–1850 Ma ago (Morey and Van Schmus, 1988; Hoffman, 1988). The Mistassini Group cannot be dated so definitely. The iron-bearing unit at the top of its succession is correlated with the Socoman Formation of the Labrador trough (Douglas, 1972) dated at 1880 Ma (U–Pb data on zircon from volcanics; Hoffman, 1988), whereas the Rb–Sr whole-rock age of the shales of the group proper (1749 ± 54 Ma; Hofmann, 1978) seems to indicate the time of a superimposed event. The volcanics of the Belcher Group of the Hudson Bay have a Pb–Pb isochron age of 1960 ± 80–1810 ± 30 Ma, and superimposed dislocations are dated here as 1800 Ma old (Hoffman, 1988, 1989).

At the very end of the Early Proterozoic, 1.8–1.65 Ga ago, the stromatolite-bearing carbonates were scarse in North America. They are known from the Hornby Bay and Et Then groups, from the Martin Formation of the northwestern Canadian shield, as well as from some other units (Grotzinger, 1988a; Ross and Donaldson, 1988). Of all the stromatolites, only *Externia externa* Semikh. and *Stratifera taenica* Semikh. from the Et Then Group were described (Semikhatov, 1978).

In the Late Proterozoic, especially in the Riphean, the American stromatolites expanded their geographic distribution considerably. Hofmann (1981) indicated 29 areas of their occurrence in Canada, and some of them show microbiolites at several stratigraphic levels. In addition, there are more than ten areas in the United States and Mexico (see Bertrand-Sarfati and Awramik, 1992 for the list). Most of the areas are located in the northwestern Canadian shield and the Rocky Mountains of Canada and the United States. Regretfully, adequate information on the taxonomic composition of the respective stromatolites is very fragmental and from only a few localities. Therefore, the available data on the Riphean and Vendian stromatolites of North America reflect our knowledge on their composition in various horizons of the sections rather than the actual diversity dynamics.

Among the Early Burzyanian stromatolites of North America, certain buildups from the Sibley Group of

the Great Lakes area are the only forms with a valid description. They were first identified as *Conophyton* cf. *garganicum* Kor. (Hofmann, 1969), and afterwards, as *Ephyaltes* sp. (Grey, 1994a). The whole-rock Rb–Sr age of shales in this group is 1339 ± 39 Ma, but the U–Pb zircon age of 1545 ± 10 Ma for volcanic rocks suggests that they are from the Lower Burzyanian interval (Davis and Sutcliffe, 1985).

The younger Proterozoic stromatolites of North America come mostly from the Belt Supergroup of Glacier National Park (USA), the Mackenzie Mountains Supergroup of northwestern Canada, and some other units of the Rocky Mountains. The age interval of the Belt Supergroup, which was based for a long time on the K–Ar and Rb–Sr measurements on glauconite and whole-rock shale samples, was recently depicted more precisely according to the paleomagnetic and more reliable isotopic-geochronologic data (see Horodyski, 1993; Link *et al.*, 1993 for review and bibliography). The supergroup rests unconformably on the crystalline rocks with a U–Pb zircon age of 1576 ± 13 Ma and older. Its basal horizons include intrusive bodies, which are 1433 ± 10 (U–Pb data on zircon) and 1304 ± 50 Ma (Rb–Sr data on whole-rock samples) old. The upper horizons of the supergroup contain glauconite with the Rb–Sr age of 1170 ± 20 Ma, and are cut by a sill with the K–Ar age of about 1100 Ma. However, paleomagnetic correlations with better dated successions suggest that the age of upper horizons of the Belt Supergroup is approximately 1250 Ma. This is used in the current literature to assure that the time range of the Belt Supergroup is 1450–1250 Ma (Link *et al.*, 1993). In other words, it corresponds to the Late Burzyanian and Early Yurmatinian.

The Burzyanian–Yurmatinian (Lower–Middle Riphean) boundary can be recognized in the supergroup succession according to stromatolite composition only. Its stromatolites are divisible into two assemblages. The following taxa can be mentioned in the lower assemblage confined to the Ravalli Group: *Colonnella frequens* (Walc.) (=*Collenia albertensis* Fent. et Fent., =*Col. columnaris* Fent. et Fent.), *Collenia undosa* Walc. and *Colleniella symmetrica* (=*Collenia clapii* Fent. et Fent.). The upper assemblage reported from higher levels of the supergroup consists of the three mentioned taxa associated with *Colleniella vermiformis* (Fent. et Fent.), *Stratifera multiflabella* (Fent. et Fent.), *Conophyton cylindricus* Masl. (=*Con. inclinatum* Rez.), *Jacutophyton* sp., *Omachtenia* sp. (=*Collenia compacta* Walc.), and, most significantly, with *Baicalia columnaris* (Fent. et Fent.), the representative of the group *Baicalia* appears in the Middle Riphean. All of the above strongly suggests that the lower and upper stromatolite assemblages of the Belt Supergroup are of Late Burzyanian and Early Yurmatinian age, respectively. It should also be mentioned that *Collenia undosa* Walc. occurs in the lower part of the Unkar Group of Arizona, which is ca. 1200–1250 Ma old (Link *et al.*, 1993).

Of the Late Yurmatinian stromatolites of North America, the following forms were described or identified as valid: *Tungussia mescalina* Bertr.-Sarf. et Awramik, *T. chrysolita* Bertr.-Sarf. et Awramik, *Apachina henryi* Bertr.-Sarf. et Awramik, and *Conophyton* cf. *garganicus* Masl. from the Mescal Limestone of the Apache Group, Arizona (Cloud and Semikhatov, 1969; Bertrand-Sarfati and Awramik, 1992); *Baicalia* sp. and *Jacutophyton* sp. from the Crystal Spring Dolomite, eastern California (Howell, 1971); and new representatives of groups *Svetliella, Baicalia, Conophyton, Jacutophyton*, and *Stratifera* from the lower part (map unit H-1) of the Mackenzie Mountains Supergroup, northwestern Canada (Aitken *et al.*, 1978b; Semikhatov, unpublished data).

The Apache Group is younger than the anorogenic granites and older than the diabases which have the U–Pb zircon ages of 1440–1420 and 1120 ± 10 Ma, respectively. The paleomagnetic correlations suggest that the age of the Mescal Limestone is close to 1150 Ma (Link *et al.*, 1993). The age of the Crystal Spring Dolomite is estimated as 1200–1150 Ma (Link *et al.*, 1993). As for the mentioned deposits of the Mackenzie Mountains Supergroup, they were assigned to the Upper Yurmatinian on the basis of the following data. The supergroup is cut by diabases with the U–Pb age of 778 ± 2 Ma (Jefferson and Parrish, 1989), and lithostratigraphic correlations (Young, 1977; Jefferson and Young, 1988), as well as drilling and seismostratigraphic records (Cook, 1992) imply that it is younger than 1270 ± 4 Ma, the age of basic volcanics of the northwestern Canadian shield. Paleomagnetic correlations suggest that the upper part of the supergroup (the Little Dal Group) is ca. 800–850 Ma old, and the middle part (the Katherine Group) is ca. 900–1000 Ma old (for bibliography see Link *et al.*, 1993). The stromatolite composition in the supergroup is in accordance with such conclusions.

The Late Riphean stromatolites have a broader distribution in North America than their Middle Riphean forms, and occasionally constitute biostromes of unique extent (up to 1200 km; Jefferson and Young, 1988). However, their taxonomy was studied to some extent in only a few areas. Among the valid taxa, the following belong to the Early Karatavian: *Inzeria tjomusi* Kryl. and *Baicalia* sp. 2 from the Katherine Group of the Mackenzie Mountains and the comparable deposits of Victoria Island (Aitken *et al.*, 1978a, 1978b; identifications by Semikhatov); *B. lacera* Semikh. from the lower part of the Tindir Group of Alaska; *Stratifera* sp. and *Tungussia occidentale* (Dow.) from the Galeros Formation, the Chuar Group, Arizona (Semikhatov, 1974); and probably *Colonnella* sp., *Jacutophyton* sp., *Conophyton* sp., and *Platella* sp. from the Gamusa Formation of northern Mexico (Weber *et al.*, 1979; Cevallos-Ferriz and Weber, 1980; Cevallos-Ferriz *et al.*, 1988) and *Archaeozoon accadiense* (Matt.) from the Green Head Group of New Brunswick (Hofmann, 1974). As follows from the published description of buildups from the mentioned and correlative units

(Hofmann, 1974; Jefferson and Young, 1988; Jackson and Jannelli, 1988; and others), this list is far from representative of the real diversity of respective stromatolites.

The age of the Katherine Group was discussed above. The Chuar Group is younger than basalts with Rb–Sr age of 1070 ± 30 Ma and, according to indirect considerations, older than 800 Ma. Paleomagnetic correlations suggest that the lower part of the group (the Galeros Formation) accumulated about 850–900 Ma ago, while the upper one (the Kwagunt Formation) is about 800–850 Ma old and correlated with the Little Dal Group of the Mackenzie Mountains (for review and bibliography see Horodyski, 1993; Link *et al.*, 1993). The Gamusa Formation and the Green Head Group are less reliably related to the time scale. The above formation is a constituent of the carbonate-siliciclastic succession, which rests unconformably on the 1100-Ma-old and older crystalline rocks, and is overlain by the Lower Cambrian deposits. According to lithostratigraphic criteria and the general position in the section, this succession is usually correlated with the Vendian deposits of eastern California (Stewart *et al.*, 1984; Link *et al.*, 1993); however, the stromatolite composition suggests the Early Karatavian age, and this age is conventionally accepted for the Gamusa Formation. The same can be said with regard to the stromatolites of the Green Head Group, whose minimal age limit is estimated to be 750 ± 80 Ma on the basis of the Rb–Sr datings of its volcanic rocks (Hofmann, 1974).

The Upper Karatavian of North America shows a slight increase in diversity of the identified taxa. These are the following forms: *Inzeria tjomusi* Kryl., *Parmites concrescens* Raab., *Minjaria uralica* Kryl., and new representatives of groups *Baicalia, Stratifera, Conophyton, Boxonia*, and *Gymnosolen* from the Little Dal Group of the Mackenzie Mountains and correlative deposits of the Victoria Island (identifications by Semikhatov); *G.* sp. 2 from the Kwagunt Formation of Arizona (Semikhatov, 1974); and *Tungussia* (=*Eleonora*) *ramosa* (Bertr.-Sarf.), *Poludia borealica* Bertr.-Sarf., *P. tyrrellina* Bertr.-Sarf., *Inzeria groenlandica* Bertr.-Sarf., and *Jurusania* sp. from the Eleonore Bay Group of eastern Greenland (Bertrand-Sarfati and Caby, 1976). As for the age constraints for the Kwagunt Formation, we should add that this formation contains *Mellanocirillium* (Horodyski, 1993), the form common in the Upper Karatavian deposits. The Eleonore Bay Group is younger than 1100 Ma and older than the Vendian tillites; its microfossils suggest Late Karatavian (700–800 Ma) age (Green *et al.*, 1989). The above age interpretation of all listed units accords well with the fact that many genera and several species of their stromatolites are typical of the Upper Karatavian of northern Eurasia.

The presented data show a general gradual increase in the diversity of the valid stromatolite forms of North America during the Riphean. The reversed trend is typical of the Vendian. The Vendian stromatolites of North America are known only in the Rocky Mountains of the United States and Canada. They constitute small bodies subordinate to siliciclastic deposits. The only forms with valid identification, among others, are *Boxonia* aff. *gracilis* Kor., *Linella* aff. *ukka* Kryl., and *Paniscollenia* sp. from the Johnnie Dolomite of eastern California (Cloud and Semikhatov, 1969). The Vendian age of the dolomite (600–650 Ma) is inferred from the stratigraphic position of the unit between the Late Karatavian tillites (about 760 Ma) and the fossiliferous Lower Cambrian deposits (Link *et al.*, 1993). There is no reliable taxonomic data on the Cambrian stromatolites of North America.

In sum, we should note the following: (1) the diversity histogram for the Proterozoic stromatolites of North America has a generally smoothed form that is related to their inadequate studies; (2) available materials show that the diversity of the Early Proterozoic stromatolite buildups was at the peak 2.0–1.8 Ga ago, and very low before and after this interval; (3) according to fragmental records on the Late Proterozoic stromatolites of North America, the diversity of their valid taxa tended to increase very slowly from the Early Burzyanian until the Late Karatavian, and reduce during the Vendian (Fig. 3); this trend is valid even if the Gamusa Formation of Mexico is assigned not to the Early Karatavian, as it was done in the present study (see above), but to the Vendian (Stewart *et al.*, 1984; Link *et al.*, 1993).

The outlined trends of diversity variations of the Early Proterozoic stromatolites of North America (Fig. 3) are similar to those of Australia (Fig. 2), and can be well correlated with the evolution of paleotectonic and depositional settings. In America, deposits of the first half of the Early Proterozoic with very low stromatolite diversity have a narrow distribution, and are dominated by immature clastic and volcanogenic successions deposited either in orogenic belts, or somewhere on the epi-Archean platforms, and sometimes (2.5–2.2 Ga ago) under the glacial climatic conditions (Hoffman, 1988, 1989; Ojakangas, 1988; Goodwin, 1991). The increase in stromatolite diversity 2.3–2.0 Ga ago, which is recorded in all Eurasian superregions (Semikhatov and Raaben, 1994, Figs. 2–4), is hardly recognized in North America at the present state of knowledge. The relative diversity peak of the American stromatolites 2.0–1.8 Ga ago was related to the appearance of carbonate shelves along passive continental margins and flanks of some orogenic belts which originated 2.0–1.92 Ga ago and terminated during the Hudsonian (Penokean) orogeny 1.86–1.82 Ga ago (Hoffman, 1989; Goodwin, 1991). The decrease in the North American stromatolite diversity (and abundance) at the end of the Early Proterozoic can be explained by the extensive uplifts that followed the Hudsonian orogeny and gave rise to the deposition of molasses and subaerial volcanism of that age. The available data on the Late Proterozoic stromatolites of North America, reflects, as it was mentioned above, the degree of our knowledge on composition of the build-

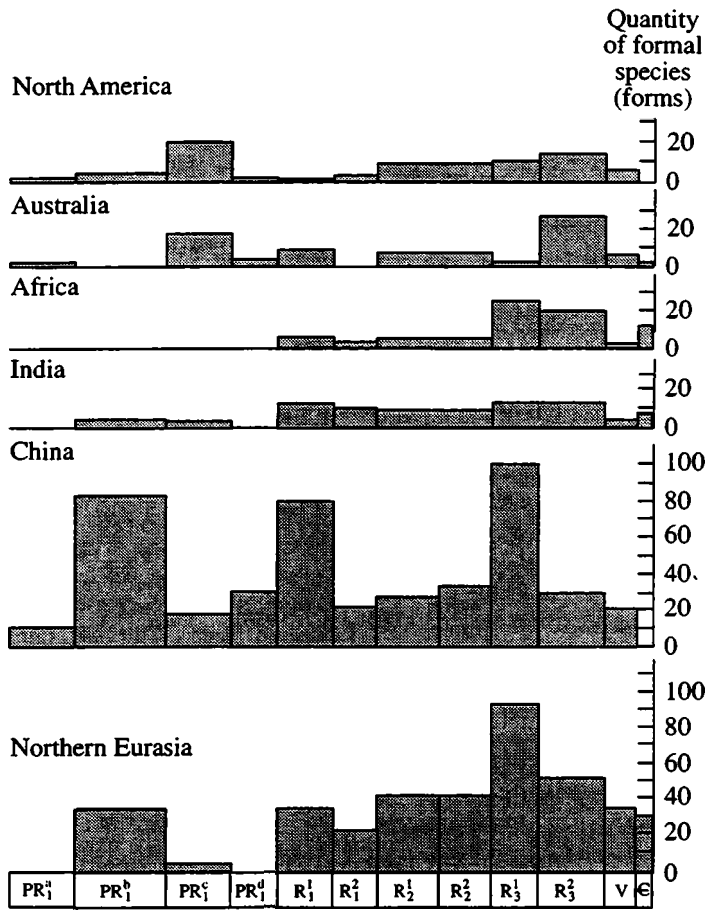

Fig. 4. Secular variations in species diversity of Proterozoic stromatolites from Northern Eurasia, China, India, Africa, Australia, and North America.

ups in different horizons of the section rather than real secular variations in diversity. Therefore, this data are not considered here in the historical-geological context.

COMPARISON OF THE SUPERREGIONAL SAMPLES AND DYNAMICS OF THE GLOBAL DIVERSITY OF PROTEROZOIC STROMATOLITES

The above and earlier published data (Semikhatov and Raaben, 1993, 1994) demonstrate that the diversity dynamics of the Proterozoic stromatolites at specific and generic levels was similar in each of the superregions under analysis and, despite some peculiarities, all the superregional diversity histograms demonstrate important common features.

Their similarity is especially distinct when the Early and Late Proterozoic intervals of the histograms undergo separate examination. The most similar are the Early Proterozoic secular changes in different-size samples from Northern Eurasia, China, and India (Fig. 4). Their main feature is a peak of variable height over the interval of 2.3–2.0 Ga, which is bounded by two deep (down to zero) depressions. However, the Chinese histogram also shows a unique notable increase in the diversity at the very end of the Early Proterozoic (1.8–1.65 Ga). The Early Proterozoic parts of two other histograms, (for Australia and North America) are similar in the peak displacement from the second to third subdivision of the Early Proterozoic (2.0–1.8 Ga), and in the very low diversity during the rest of the Early Proterozoic (Figs. 1, 3, and 4). As for the African histogram, its characteristic zero depression throughout the entire Early Proterozoic is a unique feature not typical of other superregions.

The most similar Late Proterozoic histograms are the Northern Eurasian and Chinese, the superregions with the richest taxa samples. Their common features are as follows: two diversity peaks, the lower of which is confined to the Early Burzyanian, and the higher one to the Early Karatavian intervals; the asymmetric depression between the peaks with the steep left and gentle right sides, and the rapid diversity decline during the Late Karatavian and Vendian (Fig. 4). Likewise, there is an obvious similarity between the Late Proterozoic histograms for India, Africa, and Australia (Figs. 1, 2, and 4). The distinctive feature here is a very deep and flat depression corresponding to the main part of the Riphean and separating two peaks of relative diversity. The less distinct peak marks the Early Burzy-

anian time, and the more notable one corresponds to the Early Karatavian in Africa, the Late Karatavian in Australia, and the whole Late Riphean in India according to the present state of our knowledge. At the same time, these histograms demonstrate a decrease in stromatolite diversity at the end of the Late Proterozoic, i.e., during the Late Karatavian and/or Vendian. As mentioned above, the corresponding histogram for North America is less representative, and looks specific in light of the available incomplete records. Nevertheless, the peak and subsequent decrease in the diversity occupy the same age intervals in this histogram as in the Australian one.

Thus, a majority of the discussed histograms demonstrate the trimodal pattern of higher stromatolite diversity from the beginning of the Proterozoic towards the end of the Riphean, and subsequent its significant fall off and location of minima at the following intervals: 2.3–2.0 (Northern Eurasia, China, India) or 2.0–1.8 Ga ago (Australia), the Early Burzyanian (Northern Eurasia, China, India, Australia), and then the Early (Northern Eurasia, China) or Late Karatavian (Australia), or the entire Karatavian (India). Exceptions are the American and African histograms, which are of a bimodal character. The first of them is devoid of the Early Burzyanian peak, and in the second one, the Early Proterozoic peak is missing. However, due to the positions of two other peaks, the American histogram appears to be similar to the Australian one, and the African histogram resembles the Chinese and Northern Eurasian graphs (Fig. 4). The revealed similarity between the superregional histograms suggests that each of them reflects large-scale events which are beyond superregional significance.

Figure 5 presents the summary of results for these superregional samples exemplifying continents with reasonably adequate identification of stromatolites. In essence, it is a graph of the global stromatolite diversity based on the secular trends of 236 formal genera and 908 taxa of a specific level (the procedure of counting them was discussed above). The latter value is only slightly distinct from the sum of all taxa in the superregional samples, because only about 8.5% of the species are recorded as crossing the boundary of a single superregion. The corresponding value for genera is noticeably higher, up to 31%. The low percentage of species in common between two or more superregions emphasizes the endemic character of coeval, but laterally discrete, stromatolite assemblages (Semikhatov, 1985) and, to some extent, the presence of unidentified synonyms.

All of the above suggests that the histograms in Fig. 5 adequately depict global variations in diversity of Proterozoic stromatolites at specific and generic levels. Characteristic features of the histograms are as follows: (1) the trimodal character with the peak location at 2.3–2.0 Ga, Early Burzyanian, and Early Karatavian intervals; (2) minima of taxonomic diversity at the very beginning (2.5–2.3 Ga) and very end (1.8–1.65 Ga) of the Early Proterozoic; (3) the abrupt decrease in the taxonomic diversity during the Late Burzyanian and the gradual rise during the Yurmatinian; (4) the considerable diversity drop in the Late Karatavian, which continued in a smoothed form into the Vendian. Literally interpreted, the available data implies the continuation of this drop into the Cambrian as well, but this inference is disputable due to the insufficient study of the Cambrian stromatolites.

In spite of the similarity between the global histograms for species and genera, each of them is marked by some peculiarities. For instance, the species histogram shows the Early Karatavian maximum to be 1.7 times higher than two other peaks of almost equal size, and a very wide amplitude of variations. The histogram for genera is less contrasting: the Early Karatavian peak is not so prominent, while the peak in the middle of the Early Proterozoic (2.3–2.0 Ga) is unexpectedly the highest. This resulted from the interaction of three different factors: (1) the increase of the average number of species per genus during the Proterozoic (not more than 1.7 in the Early Proterozoic and 3.2 in the Early Karatavian); (2) the higher endemism of the Early Proterozoic stromatolites in comparison with that of the Riphean and Vendian forms (Semikhatov, 1985); (3) the apparent inflation of stromatolitic genera in China, especially their branching varieties, dominate among other taxa and thus contribute much to the peak value under consideration. To put it differently, the available data do not provide the unambiguous conclusion that the morphological (=generic) stromatolite diversity was at the maximum 2.3–2.0 Ga ago. However, these data show that the diversity of microstructures, which is the main or single criterium in the modern classification of species, was at the maximum in the Early Karatavian and then declined.

Secular variations in the inner structure of stromatolitic assemblages, i.e., in ratios between representatives of different supergeneric groupings, are shown in the circular diagrams of Fig. 5. The main trends of these variations are the following: taxa of columnar branching buildups show two upward trends of relative abundance during the Proterozoic (from the beginning of the Proterozoic until 2.0–1.8 Ga and from the Middle Riphean until the Late Karatavian), which were followed by the periods of decline (1.8–1.35 Ga ago and during the Vendian); the Riphean taxa of stratiform and nodular buildups exhibit abundance variations opposite to those of columnar forms; the role of microstromatites generally increased from the beginning of the Proterozoic until the Early Burzyanian, and sharply reduced during the Late Burzyanian and especially the Middle and Late Riphean down to zero in the Vendian. As a result, each of the maxima and minima of the histograms illustrating global variations in stromatolite diversity is characterized by the specific proportions between taxa of different supergeneric groupings.

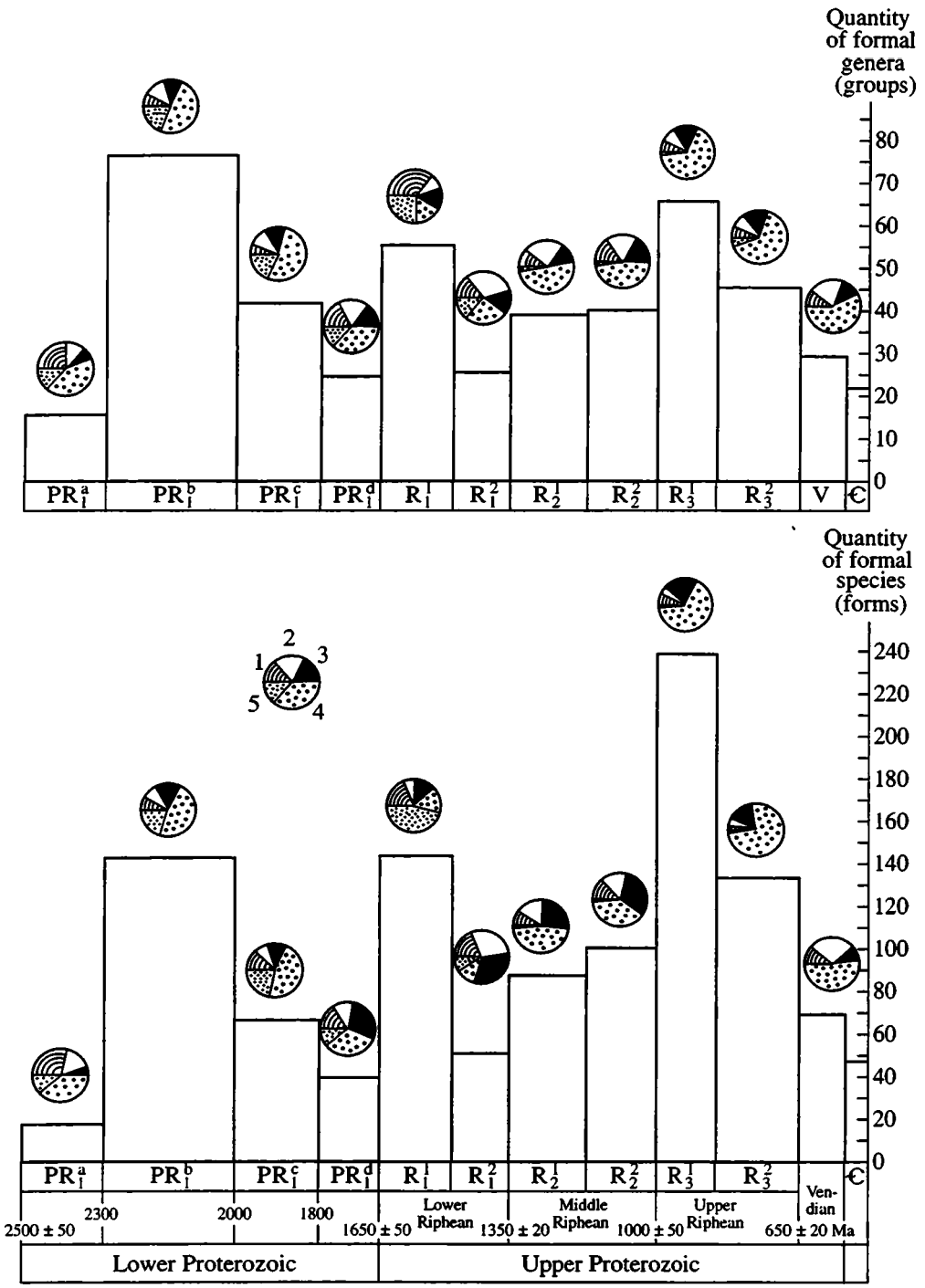

Fig. 5. Secular variations in global diversity of Proterozoic stromatolites at the levels of formal species (forms) and formal genera (groups) and secular structural changes in their assemblages

(1–5) percentages of species and genera in the following supergeneric categories of stromatolites: (1) stratiform, (2) nodular, (3) columnar unbranching, (4) columnar branching, (5) microstromatites.

CONCLUSION

The available data indicate that secular variations in the taxonomic diversity of the Proterozoic stromatolites was controlled by the multiple-factor system with a ramified network of feedbacks. The diversity was, to some extent, correlated with the stromatolite abundance, and the secular quantitative changes in the successive stromatolitic assemblages occurring against the background of their qualitive transformations—the unique succession of taxa with definite time ranges.

The most evident factor that influenced the diversity (and abundance) of the Precambrian stromatolites was the tectonic evolution with its long-term and relatively short-term trends. The first trend corresponds to the stabilization of cratonic areas, and the second trend to such processes as subsidence and elevation of these cratons, their collision, development of mobile belts and orogenesis.

The occurence of carbonate platforms and relatively diverse stromatolites of the Late Archean suggest that the biosphere of that time was favorable for an intensive stromatolite formation and considerable carbonate accumulation (Grotzinger, 1989, 1990). However, an extensive formation of carbonate platforms with inherent masses of stromatolite took place in the Early and especially Late Proterozoic, after the emergence of the epi-Archean and much larger epi-Svecofennian (epi-Hudsonian) cratons. The new tectonic situation predisposed the formation of vast and shallow epicratonic basins and gave rise to the upwelling with considerable amounts of metabolites put into circulation. Thus, as it was already noted in literature, the new ecological environments created were favorable for the procaryote development and, consequently, for the stromatolite formation.

However, the emergence of large and stable continental masses does not signify *a priori* the high stromatolite diversity. At the very beginning of the Early Proterozoic (2.5–2.3 Ga), the diversity was insignificant because of the limited distribution of supracrustal, especially carbonate deposits of that age, which partly accumulated under severe climatic conditions of the Canadian Glacioera (Chumakov, 1987). The interference of short-term tectonic factors was needed to trigger a peak of stromatolite diversity 2.3–2.0 Ga ago, as it is evident from geological records in Karelia, Scandinavia, and China. At that time, 2.2–2.1 Ga ago more precisely, specific features of the geodynamic and paleogeographic conditions in mobile belts of northern China, the Baltic shield and epi-Archean craton of Karelia provoked intensive carbonate accumulation in these areas. The paleoclimatic changes—the termination of the Canadian Glacioera and aridization in these regions—also contributed to the carbonate accumulation (for bibliography see Semikhatov and Raaben, 1994).

The sharp decrease in the global stromatolite diversity about 2.0 Ga ago can be attributed, in a tectonic context, to both the Lulang orogeny in mobile belts of China, and the reduction of carbonate accumulation in the Baltic shield. This decrease was somewhat buffered by the simultaneous enlargement of relatively small Australian and North American samples, which are related to the specific time range of carbonate platforms in these continents (Grotzinger, 1989, 1990; Gee and Grey, 1993). Nevertheless, the end of the Early Proterozoic, especially 1.8–1.65 Ga ago, is marked by a rapid drop in the diversity and abundance of stromatolites. No doubt, this was a consequence of significant tectonic events, such as the Svecofennian, Hudsonian and Barramundian orogeneses, which almost simultaneously affected (1.92–1.80 Ga ago) the vast territories of Northern Eurasia, North America, and Australia (Hoffman, 1988, 1989; Goodwin, 1991; Gorbatschev and Bogdanova, 1993; Rosen *et al.*, 1994), and caused the elevation of several cratons, a wide distribution of clastic and volcaniclastic complexes, including terrestrial deposits, and, correspondingly, a sharp reduction of carbonate sedimentation (Semikhatov, 1974; Goodwin, 1991). The latter was still under progress at the end of the Early Proterozoic only in the epi-Lulang craton of northern China. The laterally restricted younger orogenesis of the pre-Riphean time was also responsible for these processes (Goodwin, 1991; Gorbatschev and Bogdanova, 1993).

The Early Burzyanian peak of stromatolite diversity in the global (Fig. 5) and many superregional histograms (Fig. 4) is correlative with the extensive marine transgression, which led to the expansion (northern China) or origination of large carbonate platforms on continents, and to the formation of some mobile belts (e.g., the Kibaran belt) with noticeable amounts of carbonate deposits. The considerable global decrease in stromatolite diversity at the end of the Early Riphean is mostly induced by the largest samples from Northern Eurasia and China. In China, the diversity decline coincides with the further development of marine transgression commenced at the end of the Early Proterozoic (at the early Changchen time), while in Northern Eurasia and Australia, it was during the period of general regression and low sea level stand. The contrasting character of events suggests another, but not paleogeographic cause of the Late Burzyanian stromatolite decline.

The global stromatolite diversity was growing from the Early Yurmatinian to the Early Karatavian time against the background of two main geological events: (1) the progressive subsidence of stable blocks with corresponding expansion of transgression and carbonate sedimentation that was especially typical of the Early Karatavian time in Africa, eastern China, and Middle Siberia; (2) a formation of a series of intracratonic mobile belts, the largest of which appeared about 1.0 Ga ago in Africa (Semikhatov, 1974; Goodwin, 1991; Trompette, 1994). However, these events were hardly the main and sole cause of the Early Karatavian diversity peak: the scale of the maximum expansion of paleobasins and carbonate platforms does not coincide with that of the stromatolite diversity, and the mentioned peak occured against the background of other significant biospheric events as it is outlined below.

The Late Karatavian drop in stromatolite diversity seems to be directly influenced by the following contemporaneous events: (1) almost synchronous (850–800 Ma) Early Baikalian, Jinning and Lusakan orogeneses in several mobile belts, and the wide distribution of siliciclas-

tic formations in China, southwestern and Equatorial Africa, and some Siberian areas; (2) the emergence of extensive stable massifs in Northern Eurasia and northern China during the Late Karatavian. The relevant decrease in the stromatolite diversity detectable in Northern Eurasia, China, and Africa was not compensated at the global scale by the synchronous diversity rise in Australia and North America, where the latter was a response to the specific development of particular tectonic structures.

As for the further Vendian drop in stromatolite diversity, it was coeval with contrasting tectonic events, which should produce opposite effects on stromatolite development. On the one hand, these were the pan-African events which resulted in a wide molasse distribution in the Gondwanan continents and influenced the balance between continental and mantle fluxes into oceans (see below). On the other hand, it was the postglacial Vendian transgression, when the enormous mass of shallow water carbonates potentially favorable for stromatolite development, though deficient in them, were accumulated in Siberia, southern China, and some other regions.

Some correlations between stromatolite diversity and geodynamics are inferable from secular variations of $^{87}Sr/^{86}Sr$ ratio in sea water, which are known to be under the control of the global balance between the two main sources of material entering oceans: the continental runoff and mantle flux (for review and bibliography see Kaufman *et al.*, 1993; Gorokhov *et al.*, 1995). Generally, the ascending Riphean and Vendian variation curve for this ratio, which reflects the Sr isotopic evolution in provenance and the increasing volume of eroded continental crust, demonstrates a noticeable drop at the end of the Late Yurmatinian and in the Early Karatavian time. This drop was caused by the less intense continental erosion in response to the termination of the Grenvill events and the advancing Early Karatavian marine transgression, as well as by the sharply increased hydrothermal flux from the mantle probably related to the considerable intensification of riftogenesis and changes in configuration of divergent plate boundaries. Thus, the Early Karatavian peak of stromatolite diversity (and abundance) was formed in conditions of the high-intensity hydrothermal flux into the World Ocean.

The Late Karatavian drop in diversity was coeval with an increase of $^{87}Sr/^{86}Sr$ ratio in sea water presumably related to the Jinning–Early Baikalian events and the extensive glaciation of the African Glacioera (Chumakov, 1987). The latter could not but cause a significant sea level drop and, hence, an emergence of shelves and their erosion. The Vendian diversity decrease was under conditions of a rapid increase of the radiogenic ^{87}Sr input into the World Ocean, which was unique in the Precambrian history of the earth. It was in response to the Pan-African orogenic processes comparable in scale with the Himalayan orogenesis, and was not compensated by a simultaneous growth of the mantle flux (for details see Kaufman *et al.*, 1993; Gorokhov *et al.*, 1995).

The diversity and abundance of Proterozoic stromatolites were unique in the earth's history, and their dynamics reflects not only the changes in dominant geodynamic and paleogeographic conditions, but also chemical properties of the medium. It was shown recently that the Precambrian and especially Early Proterozoic sea water was oversaturated with respect to Ca, and the ratio of $(HCO_3)^-$ to Ca^{2+} might be higher than in the Phanerozoic sea water (Grotzinger, 1989, 1990; Grotzinger and Kasting, 1993). As a result, the Early Proterozoic and, to less extent, the Riphean and Vendian were periods of significant precipitation of the calcite and aragonite marine cement in normal marine settings, while the absorption of CO_2 by microbial mats led to the abundant accumulation of aragonite and (high-Mg?) calcite within these mats. This model chemically substantiated the empirical inference, which has long been developed by Russian scientists, who suggested that the Precambrian stromatolites were mainly formed by means of biologically induced carbonate sedimentation leading to syngenetic lithification of buildups (bibliography see Serebryakov, 1975).

Thus, the stromatolite abundance is highly dependent on the most important parameters of carbonate accumulation, such as the degree of sea water saturation with respect to Ca and Mg carbonates, the CO_2 regime in the hydrosphere, the earth's surface temperature, etc. (Grotzinger, 1989, 1990; Sochava and Podkovyrov, 1992). In this respect, the narrow time range of the abundant occurrence of microstromatites (*Asperia*-form buildups) attracts special attention (Raaben, 1980; Grey and Thorne, 1985; Grotzinger, 1989). As it was detected, microstromatites played an important role in the pre-Late Burzyanian stromatolite assemblages, and disappeared from geological records almost completely during the Middle Riphean (Fig. 5, circular diagrams). They were important attributes of tidal flats, and are regarded now, completely or partly, as abiogenic structures, carbonate tuffs, or precipitates. Therefore, their decline may indicate a fall in both sea water saturation with carbonates and chemogenic carbonate precipitation in the extreme shallow water settings (Grotzinger, 1989, 1990; Grotzinger and Kasting, 1993). Following Grotzinger (1989, 1990) and Sochava and Podkovyrov (1992, 1993), we can assume that the Early Karatavian peak of species diversity of stromatolites corresponded to an optimal degree of sea water saturation with respect to Ca- and Mg-carbonates. In addition to boitic factors, these conditions favored the maximal diversity of stromatolite microstructures, but the saturation degree was not high enough to induce an intense chemogenic carbonate sedimentation that could suppress the biologically induced precipitation. This inference seems to be tempting in light of the recent data on the evolution of Proterozoic carbonate accumulation and specific lithological, mineralogical

and chemical peculiarities of the Late Riphean (or Late Karatavian) carbonates as compared with their younger and older counterparts (Grotzinger, 1989, 1990; Fairchild *et al.*, 1990; Knoll and Swett, 1990; Sochava and Podkovyrov, 1992, 1993; Knoll and Sergeev, 1995).

As for the Late Karatavian decline in stromatolite diversity, it corresponds to the time range of the African Glacioera (Chumakov, 1978, 1987). Although the African materials do not show the direct effect of the first Late Riphean glaciation epoch on the composition of regional stromatolite assemblages, the expansion of the Late Karatavian and Early Vendian glaciations implies a noticeable decrease in the atmospheric pCO_2 (Kasting, 1987). Apparently, the low value of pCO_2 changed the degree of sea water saturation with Ca-carbonates that, in turn, affected the intensity of biogenic carbonate precipitation inside or near the cyanobacterial mats (Knoll and Swett, 1990; Semikhatov and Raaben, 1993).

Finally, the Vendian decrease in stromatolite diversity occurred after the termination of the Laplandian (Varangerian) glaciation, the most considerable in the earth's history, against the background of a changing Ca/Mg ratio in sea water, pCO_2 growth, and an increasing earth surface temperature due to the greenhouse effect (Brasier, 1992). These and the above discussed changes in the Vendian paleogeographic environments produced the doubtless favourable prerequisites for the increase in stromatolite abundance and diversity. The evident contradiction between these prerequisites and the real data suggests that for the Vendian time we have to seek of another powerful factor influencing the intensity of stromatolite formation.

Returning to the Early Proterozoic peak of stromatolite diversity, we remind that it first of all reflects data from the Upper Jatulian of the Baltic shield and the middle part of the Hutuo Supergroup of China, whose deposits accumulated ca. 2.2–2.1 Ga ago (Semikhatov and Raaben, 1994). In the Baltic shield and some other regions, the carbonates of this age were shown to have very high values of $\delta^{13}C$ ($10 \pm 3‰$), which reduced abruptly (>10‰) 2.1–2.06 Ga ago (Karhu, 1993; Tikhomirova and Makarikhin, 1993). Taking into account mechanisms controlling the isotopic composition of the carbonate carbon (the relationship between the burial rates of the carbonate and organic carbon and the degree of isotopic fractionation in the biologic carbon cycle), we consider the mentioned $\delta^{13}C$ values as an indicator of the increased bioproductivity of organic photosynthesis and higher intensity of C_{org} burial 2.2–2.1 Ga ago, and their subsequent decrease. It is logical to think that these events caused the appearance (or increased content) of the atmospheric oxygen and oxidation conditions in the upper layer of the World Ocean about 2.0 Ga ago (Kasting, 1987; Cloud, 1980; Kasting *et al.*, 1992; Karhu, 1993; for a different point of view see Tikhomirova and Makarikhin, 1993), thus being responsible for the most significant changes in the biosphere.

At the end of the Early Proterozoic and during the Early and Middle Riphean time, the stromatolite diversity variations were not accompanied by variations between the burial rates of oxidized and reduced carbon: $\delta^{13}C$ values in carbonates of this age were in the range of $0 \pm 2‰$ and tended to grow at the very end of the Middle Riphean only (for bibliography see Knoll *et al.*, 1995). Thus, unlike the Early Proterozoic peak of diversity, that Early Burzyanian peak was not reflected in the isotopic composition of the carbonate carbon. This can be explained either by the volume difference of relevant reservoirs of carbon, or (if the abiogenic origin of microstromatites is accepted) by the great importance of these buildups in producing the Early Burzyanian peak.

In contrast, the Late Riphean and Vendian changes in stromatolite taxonomic composition proceeded against the background of the considerably variable isotopic composition of carbonate carbon (Kaufman *et al.*, 1993; Kaufman and Knoll, 1995; Knoll *et al.*, 1995). The Lower Karatavian carbonates demonstrate low positive values of $\delta^{13}C$ (maximum about 4‰) that could be an effect of a higher rate of organic photosynthesis. This assumption agrees well with the fact that significant evolutionary changes in microbiota are recorded at the Middle–Late Riphean transition (see below), and that the Early Karatavian stromatolites are very diverse and abundant. However, the intense hydrothermal flux into the Early Karatavian oceans (see above) suggests that the oxidation potential of sea water was lowered to some extent at that time, and this alone should increase the burial of C_{org}.

In the light of these data, it seems surprising that the Late Karatavian and Vendian stromatolite decline coincided with a generally noticeable (up to 5 to 8‰) increase of $\delta^{13}C$ values in carbonates, complicated by short-range and abrupt negative (–4 to –2‰) excursions of this parameter. These excursions demonstrate chronological and cause-and-effect relations with the glacial events (Kaufman and Knoll, 1995). As for the general growth of $\delta^{13}C$ in the Upper Karatavian and Vendian carbonates, it can be logically related to the wide distribution of seaweeds and other lower eucaryots during the Late Karatavian–Vendian time (Knoll and Swett, 1990; Knoll and Sergeev, 1995), which contributed to the higher bios productivity and, hence, to the mass of buried C_{org}. It is remarkable that the productivity raising was not only relative, but absolute as well, because the Sr-isotopic data suggest much greater rates of continental erosion during the Late Karatavian–Vendian time (see above). Judging by the stromatolite decline of that time, the increased productivity of organic photosynthesis was caused not by cyanobacteria (the time of their domination had passed), but by higher groups of organisms.

Seeking to understand the relations between diversity variations of stromatolites and changes in the composition of stromatolite-forming microbiotas, we meet three obvious obstacles: (1) the extreme evolutionary conservatism of cyanobacteria, whose main stromatolite-forming groups already existed in the middle of the Early Proterozoic (Gebelein, 1976); (2) the uncertain relations between the taxonomic composition of stromatolite-forming communities and the diagnostic features of lower stromatolite taxa; and (3) the effect of diagenetic and later processes on microstructural features which serve as the basis for the identification of stromatolite species. Nevertheless, the microstructure and microlaminar shape (microfabric) of organo-sedimentary buildups (stromatolites s. str.) are commonly considered to be predominantly dependent upon biological properties of stromatolite-forming mats and, therefore, the taxa identified according to their microstructure can shed some light upon the time–space distribution of various microbial communities (Semikhatov *et al.*, 1979; Semikhatov and Komar, 1989; Awramik, 1992; Walter *et al.*, 1992; Semikhatov and Raaben, 1993).

Ergo, an inference on the general similarity between curves illustrating secular variations in the diversity of stromatolites (this study) and benthic microfossils (Schopf, 1992) during the Proterozoic seems to be very promising at first glance. However, data on microfossil composition, which were used to substantiate this inference, were published 8–10 years ago and, as Knoll noted (1994), are inadequate in regard to the recent understanding of the Proterozoic microorganism diversity.

At present, we can state that the sharp rise of stromatolite specific (and generic) diversity at the Middle–Upper Riphean transition (Fig. 5) coincided with profound changes in the Late Proterozoic microbios, such as the rise in diversity of eucaryots (green, red and, probably, chromophyte algae), the appearance of some new genera of cyanobacteria, and the redistribution of the actual roles of some preexisting genera (Knoll, 1994; Knoll and Sergeev, 1995). It is important in this context that the Early Karatavian time was marked by both the growing diversity of stromatolite microstructures and the appearance of a new type (*Fibrosida*) known in Eurasia, Africa, and North America (Semikhatov and Komar, 1989; Komar, 1989). All the data strongly suggests that the Early Karatavian growth in stromatolite diversity reflected not only the environmental changes discussed above, but also important biotic events. The appearance of eucaryots and some new groups (or size classes) of procaryots in benthic communities might expand the adaptive abilities of the latter to environmental changes and gave rise to diversity of stromatolite microstructures (Awramik, 1992; Semikhatov and Raaben, 1993).

Biotic causes also influenced the stromatolite decline at the end of the Proterozoic. Its first step, the Late Karatavian drop in diversity and abundance, might be somehow related to the taxonomic diversification and expansion of seaweeds. The latter events, detectable from direct paleontological records (Schopf, 1992; Knoll, 1994) and the interpretation of $\delta^{13}C$ variations in carbonates (see above), occurred about 900–850 Ma ago, and might result in the reduction of cyanobacterial mat distribution in the shallow-water environments and related decline of stromatolite formation (Monty, 1973; Knoll and Swett, 1990; Knoll and Sergeev, 1995). In the Late Karatavian time, the latter effect was promoted by other factors, such as paleogeographic and geodynamic changes and a decline of phytosynthesis efficiency caused by decline in pCO_2 and temperature during the African Glacioera. The destabilization of benthic cyanobacterial ecosystems due to the repeated glacioeustatic sea level oscillations also probably contributed to the Late Karatavian stromatolite decline. Considerable regressions were likely to disturb these ecosystems (Gebelein, 1976), and their reconstruction during postglacial transgressions was apparently hampered by the competition with more advanced organisms (the seaweeds in this case).

When applied to Metazoa, the latter factor was also considered to be among the causes of the Vendian stromatolite decline (Sokolov and Fedonkin, 1986; Semikhatov and Komar, 1989; Semikhatov and Raaben, 1993). However, the main factor of this decline is usually believed to be the grazing and burrowing activity of the macroscopic animals that appeared, which expelled the cyanobacterial mats into unfavorable ecological niches (Awramik, 1971, 1992; Walter and Heys, 1985; Walter *et al.*, 1982; Semikhatov and Raaben, 1993; and others). It is also possible to speculate that one of the causes of the Vendian stromatolite decline may be the competition for nutritions with eucaryots and the rearrangement of trophic connections in the ecosystems—the creation of trophic pyramid (Fedonkin, 1992). The negative influence of the above biotic causes on the Vendian stromatolites was not compensated by the action of abiotic factors favorable for their production, i.e., higher pCO_2, climatic warming, and the extensive shallow water carbonate accumulation.

Thus, the taxonomic diversity dynamics of Proterozoic stromatolites was dependent upon a complex of interrelated and independent biotic and abiotic factors, and reflected the general evolution of Proterozoic ecosystems. Since the correlation between variations of stromatolite diversity and biotic events seems to be the case for the Late Riphean–Vendian part of the geological records, we can suggest some elements of such a correlation for the earlier history. The obvious contradiction between this suggestion and the evolutionary conservatism of stromatolite-forming cyanobacteria could be resolved to a great extent by the fact that the stromatolite-forming mats evolved as communities, and their inner structural changes, dependent on the relative proportions of different taxa, might (or must) affect the processes of stromatolite formation.

At the same time, the wide expansions of stromatolitic cyanobacterial communities and the relevant rises of photosynthetic bioproductivity were, by themselves, very significant biotic events, which directly affected the intensity of biologically induced carbonate accumulation, the balance of carbonate and organic carbon, the free-oxygen content in the hydro- and atmospheres, the burial of a great volume of CO_2, and the degree of the greenhouse effect. Among these expansion periods, the Early Proterozoic (2.3–2.0 Ga ago) and, probably, the Early Karatavian ones, were the most important for the subsequent evolution of the biospheric processes. From our standpoint, the decline of stromatolite diversity (and abundance) at about 850 Ma ago, i.e., at the Early–Late Karatavian transition, is of special importance. More and more recent data suggest that this was a time of very significant rearrangement of ecosystems (Semikhatov and Raaben, 1993, 1994; Kaufman *et al.*, 1993; Knoll, 1994; Knoll and Sergeev, 1995). The above minimum was one of the pronounced manifestations of this rearrangement.

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Emsian and Eifelian Reference Sections of Mongolia

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Abstract—Results of joint research on macrofauna and conodonts are presented in the paper. Based on new data, the relationship between conodont and regional zones was established for the first time, and boundaries between the stages under consideration were refined. The Mongolian sequences were correlated with reference sections of the Salair. The available data made possible wide interregional correlations of Mongolian sections characterized by endemic fauna.

Key words: *Emsian, Eifelian, faunal assemblages, correlation*

The Devonian of Mongolia fills a peculiar place in Central Asia. Here, within large latitudinally oriented areas, the sediments of variable facies composition that yield completely different fauna occur. The Lower Devonian terrigenous deposits developed in the vast area of northern Mongolia contain a highly endemic Mongol-Okhotsk fauna. This fauna is also characteristic of the Lower Devonian sequences in Transbaikalia, the Far East, and northeast China. In contrast to those regions in Mongolia, a narrow belt of terrigenous-carbonate sediments of the same age border from the south the area of terrigenous sequences. The representatives of the Altai-Sayan, mainly Salair, fauna dominate the benthic fossil assemblages in carbonate rocks. Brachiopods, tabulates, and bryozoans have been described in monographs (Alekseeva *et al.*, 1981; Sharkova, 1981; Kopaevich, 1984), in the collection of articles *Novye taksony iskopaemykh bespozvonochnykh Mongolii* (1992), and in a number of papers.

The first biostratigraphic schemes were elaborated separately for terrigenous and carbonate sections because of their incompatibility. Stratigraphic subdivisions corresponding to the Tom'-Chumysh, Malyi Bachat, and Shandyn horizons of Salair were distinguished in carbonate sections according to brachiopod composition. These levels formed a basis for the first Devonian stratigraphic schemes of Mongolia. Well-studied Devonian sections of the Salair, correlative with the standard scale, made it possible to define stage volumes in carbonate sections of Mongolia. In terrigenous sections with the Mongol-Okhotsk fauna, this problem was solved by correlation with carbonate sequences according to reference levels corresponding to certain geological events that occurred in the basin (Alekseeva, 1990, 1993). The establishment of the volumes and boundaries of stages of the standard scale in the terrigenous sediments of Mongolia is also important for the adjacent territories, where the North American regional stages were applied until recently. The latter stages were distinguished because of the erroneous identification of homeomorphic Asian brachiopods as North American ones.

In the unified Devonian stratigraphic scheme of Mongolia, the studied sediments are included in the Chulun and Tsagan-Khalgin horizons, the Gobi Altai Formation. The most complete, and the richest in fossils, sections of the above-mentioned subdivisions occur on the southern slope of the Gobian Altai, 10 km southeastward from the Somon Shine-Dzhinst in the Tsakhirin-Khuduk area. Here, the rocks of the Pragian to Givetian stages are exposed in a single section in a normal succession. The Emsian sediments considered in this paper conformably overlie the conglomerate sequence of the Tsakhirin Formation, which is referred to the Pragian Stage.

In describing sections of the studied sequence, we used the following previously distinguished stratigraphic subdivisions (upward from the base): the Chulun Horizon and its *Uncinulus tsakhirinicus* and *Spinatrypa (Spinatrypa) galinae–Cyrtinopsis nalivkini* regional zones, the Tsagan-Khalginian Horizon and its *Leptodontella zmeinogorskiana* and *Leptagonia reticulata* regional zones, and the Gobi Altai Formation. Faunal lists are presented according to the results of the simultaneous sampling of macrofossils (tetracorals, bryozoans, and brachiopods) together with conodonts in the "sample-to-sample" correspondence. Species of other groups of organisms are not given due to inadequately exact correlations with our data. The exception is represented by the lower part of the Chulun Horizon, where T.T. Sharkova (1981) cited the list of tabulates for the thoroughly studied part of the section. Other faunal groups identified are the following: tetracorals by L.M. Ulitina; brachiopods by R.E. Alekseeva, O.A. Erlanger, G.A. Afanas'eva, and B. Mendbayar;

bryozoans by R.V. Goryunova; and conodonts by V.A. Aristov.

The sequences described below are the type sections of the Chulun and Tsagan-Khalgin horizons with corresponding regional zones.

CHULUN HORIZON

The Uncinulus tsakhirinicus Regional Zone

(1) Basal layers (5 m thick) of calcareous, dark-gray to greenish-gray argillites and calcareous reddish platy sandstones with thin limestone beds are overlaid by dark-gray flaggy limestone, interlayered by organogenic (coral) limestone. The sediments contain tabulates *Favosites kovschovi mongolicus* Schark., *Squameofavosites kulkovi* Dub., *Thamnopora compacta minima* Schark., *Striatopora abnormis* (Mir.), *Pachypora tortuosa* Schark., *Crassialveolites abramovi* Dub., *Caliapora elegans* Yanet, *Oculipora angulata* Kim, *Coenites dunginensis* Schark., tetracorals *Embolophyllum aggregatum aggregatum* (Hill), *Martinophyllum latum* (Hill), *"Mycophyllum" difficile* Zhelt., and abundant stromatoporoid colonies. The unit is 25 m thick.

(2) The above rocks grade into dark-gray and greenish, shaly calcareous argillites and siltstones. Greenish-gray tuffites (2 m thick) occur in the upper part of the unit. Bryozoans *Fistulipora catena* Kop. were found. The thickness is 10 m.

(3) Dark-gray, detrital, platy limestones with coral beds bearing tabulates and tetracorals *Embolophyllum aggregatum aggregatum* (Hill) as well as stromatoporoids. The thickness is 25 m.

(4) Calcareous, greenish-gray, coarse-grained to gravel sandstones. The thickness is 2 m.

(5) Dark-gray, detrital, platy and flaggy limestones. Abundant fossil remains are represented by the following: tabulates *Thamnopora compacta minima* Schark. and *Crassialveolites abramovi* Dub.; tetracorals *Synaptophyllum* sp. nov.; brachiopods ? *Reeftonia* sp., *Biernatium asiaticum* Aleks., *Leptagonia zlichovensis* Havl., *Ralia primigenia* Laz., *Uncinulus tsakhirinicus* O. Erl., *Stenorhynchia* sp., *Spinatrypina* (*Spinatrypina*) *bachatica* Aleks., *Atryparia gobica* Aleks., *Howellella* sp., *Cyrtina kazi* Havl., and *Cyrtinopsis sursureni* Aleks.; and conodonts *Icriodus altaicus* Aks., *I.* sp. The thickness is 18 m.

(6) Dark-gray and greenish, shaly argillites with rare beds of dark-gray limestones, alternating with organogenic dark-gray limestone. The sediments yield the following: tetracorals *Lekanophyllum* sp. nov., *Pseudomicroplasma* sp., *Embolophyllum pestereviensis* (Zhmaev), *Lyrielasma* cf. *chapmani* Pedder, *Tryplasma aequalis* Lonsd., and "*Mycophyllum*" *eifeliense* Zhelt; bryozoans *Semicoscinium* sp.; and abundant tabulates and stromatoporoids. The thickness is 32 m.

(7) Calcareous siltstones and grayish-green platy sandstones. Extrusive beds occur in the upper part of the unit. Tetracorals *Pseudonucroplasma* sp. were encountered. The thickness is 15 m.

The total thickness of the *U. tsakhirinicus* regional zone is 127 m.

The Spinatrypa (Spinatrypa) galinae–Cyrtinopsis nalivkini Regional Zone

(8) A unit of dark-gray flaggy limestones with rare thin beds of nodular limestone. Large (20–25 cm in diameter) stromatoporoid colonies, tabulates, and bryozoans are abundant. The tabulate *Alveolites levis grandis* Schark. and tetracoral *Embolophyllum aggregatum cracente* Pedder were identified. The thickness is 40 m.

(9) Alternating dark-green and reddish tuffstones. Coarse-grained to gravel sandstones prevail at the top of the unit. The thickness is 14 m.

10. Dark-gray, sometimes almost black, flaggy limestones with rare interlayers of platy limestone. Fossil remains are represented by tetracorals *Embolophyllum fasciculus* (Zhmaev), *E. aggregatum cracente* Pedder, *Glossophyllum adatkanicum* (Tcherep.), and *Thamnophyllum tabulatum* Bulv. The thickness is 30 m.

(11) Two beds of gray, vitric tuffites, alternating with dark-green, highly weathered tuffstones. The thickness is 10 m.

(12) The lower half of the unit is composed of dark-gray, flaggy limestones, with prevailing biohermal interlayers; the upper part consists of dark-gray argillaceous limestones; platy, nodular, detrital, alternating with dark-gray leafy argillites and calcareous sandstones. The sediments contain tetracorals *Tabulophyllum chernyshevi* Bulv., *Tryplasma* aff. *devonianum* (Soshk.), and *Charactophyllum shandiense* (Zhelt.); bryozoans *Fistulipora* sp. and *Semicoscinium* sp.; and rare tabulates. The thickness is 50 m.

(13) Dark-green and gray tuffstones. Lava flows with large plagioclase crystals occur in the upper part of the unit. The thickness is 15 m.

(14) Alternation of dark-gray, leafy argillites with crinoidal limestones and rare calcareous sandstone beds. Fossil remains are scarce and poorly preserved. Conodonts *Polygnathus inversus* Klapp. et Johns. and *P. perbonus* (Phill.) were found. The thickness is 15 m.

15. Dark-gray, flaggy, nodular limestones with thin uneven silicified interlayers. Crinoidal beds with crinoid segments and large stems prevail. The fauna is represented by the following: rare tabulates and tetracorals *Amplexocarinia* sp. and ?*Aphyllum* sp.; abundant bryozoans *Fistuliphragmides gobiensis* Gorjunova (gen. et sp. nov.), *Quadrisemicoscinium disereta* (Prantl.), *Fistulipora* sp., *F. falcata* Nekh., *Lioclema implicatum* Kop., and *Amurodictya mariculata* (Kop.); and brachiopods *Levenea orientalis* Aleks., *Leviconchidiella marcovae* Aleks., *Leptaenopyxis bouei* (Barr.), *Latonotoechia* cf. *latona* (Barr.), *Spinatrypa* (*Spinatrypa*) *galinae* Aleks., *S.* (*Isospinatrypa*) *subspinosa* (Laz.), *Spinatrypina* (*Spinatrypina*) *bachatica*

Aleks., *Carinatina arimaspa* (Eichw.), *Eospirifer* ex gr. *karmanovi* Khod., *Quadrythyris* sp., and *Cyrtinopsis nalivkini* Rzhon. The thickness is 20 m.

(16) A unit of dark-green and gray tuffstones. The thickness is 32 m.

(17) Dark-gray, shaly siltstones, olive-colored on the weathering surface. The thickness is 26 m.

(18) Alternation of sandstones and siltstones, dark-gray, greenish, more rarely yellowish in color. Limestone lenses yielding abundant brachiopods *Levenea orientalis* Aleks., *Leptaenopyxis bouei* (Barr.), *Coelospirella dongbejensis* Su, and *?Fallaxispirifer korovini* (Khalf.) occur in the upper part of the unit. The thickness is 104 m.

The total thickness of the *S.* (*S.*) *galinae–C. nalivkini* regional zone is 341 m.

The total thickness of the Chulun Horizon is 468 m.

TSAGAN-KHALGIN HORIZON

The Leptodontella zmeinogorskiana Regional Zone

(19) Calcareous, dark-gray conglomerate, with sandy limestone as a cement, comprising not more than 20–30% of the total rock. Pebbles and clasts are poorly sorted, from a few centimeters to 1 m in diameter. Angular and semirounded clasts and pebbles of dark-gray limestones prevail. Rounded pebbles of almost black sparkling limestones and greenish siltstones and argillites are rare. Blocks and large pebbles of dark-gray limestones yield tabulates *Squameofavosites dubatolovi* Mir., *Thamnopora compacta minima* Schark., and *Oculipora angulata* Kim., characteristic of the base of the Chulun Horizon. Brachiopods are represented by forms encountered at the top of Chulun beds: *Leptaenopyxis bouei* (Barr.), *Leviconchidiella* sp., and Atrypinae. At seven meters above the base of unit 19, conglomerates grade into greenish-brown gravelstones containing green leafy argillite beds. The thickness is 8 m.

(20) Brown, calcareous, coarse-grained sandstones interlayered with dark-gray, flaggy limestones. The whole unit is brown-colored and contains abundant tabulates *Pachyfavosites bijensis* (Sok.); bryozoans *Fistuliphragmides gobiensis* Gorjunova (gen. et sp. nov.) and *Medisemicoscinium perelegans* Plam.; and brachiopods *Discomiorthis serotinae* Aleks., *Leptodontella zmeinogorskiana* (Peetz), *L. acuta* Khalf., *Rotundostrophia magna* (Khalf.), *Eodevonaria bortegensis* Afan., *Pseudouncinulus mongolicus* O. Erl., *Coelospirella dongbeiensis* Su, and *Elythina salairica* Rzhon. The thickness is 6 m.

(21) A unit of dark-gray (brown on the weathering surface), sandy, flaggy limestones occurring on the southern steep slope of the ridge and yielding bryozoans *Laxifenestella capillosa* (Počta), and *Reteporina petalata* (Počta) and conodonts *Polygnathus costatus patulus* Klapp., *Icriodus angustus* Stew. et Sweet, *I. corniger corniger* Witt., *I. corniger leptus* Wedd., *I. retrodepressus* Bult., and *I.* ex gr. *trojani* Johns. et Klapp. The thickness is 40 m.

(22) Dark-gray, flaggy, nodular limestone, with the following: brachiopod coquina; tetracorals *Tabulophyllum chernyshevi* Bulv., *T. schlüteri* (Peetz), *Heliophyllum* aff. *halli* Edw. et Haime, *Zonophyllum* sp., and *Pseudomicroplasma* sp. nov.; bryozoans *Laxifenestella capillosa* (Počta), *Amurodictya* sp., and *Nematopora* sp.; brachiopods *Platyorthis plana* Aleks., *Discomiorthis serotinae* Aleks., *Schizophoria* sp., *Leptagonia* sp. nov., *Leptaenopyxis bouei* (Barr.), *Megastrophiella uralensis* (Vern.), *Protocymostrophia patersoni* (Hall), *Leptodontella zmeinogorskiana* (Peetz), *L. acuta* Khalf., *Mesodouvillina birmanica* (Reed), *Xystostrophia umbraculum* (Schloth.), *Caucasiproductus* sp. nov., *Leviconchidiella* cf. *mirabilis* (Rzhon.), *Uncinulus shinensis* O. Erl., *Pseudouncinulus mongolicus* O. Erl., *Areella barunica* O. Erl., *Eucharitina dobrovi* O. Erl., *Spinatrypa variaspina* Cooper, S. (*Isospinatrypa*) *subspinosa* (Laz.), *Spinatrypina* (*Spinatrypina*) *bachatica* Aleks., *Macropleura* (?) sp., *Delthyris* (*Quadrifarius*) sp., *Howellella subgregaria* (Rzhon.), *Elythina salairica* Rzhon., and *Undispirifer frequens* (Bubl.), *Cyrtina* sp. nov.; tentaculitids *Lonchidium gobiense* Dorod. et Klish.; and conodonts *Icriodus angustus* Stew. et Sweet and *I.* ex gr. *trojani* Johns. et Klapp. The thickness is 30 m.

(23) Gray limestones medium-bedded, crinoidal, with conodonts *Icriodus corniger corniger* Witt. and *I.* ex gr. *trojani* Johns. et Klapp. The thickness is 37 m.

(24) Rose-colored, crinoidal limestones with rare bryozoans and poorly preserved brachiopods. The unit contains interlayers with flint nodules. Tetracoral *Zonophyllum* sp. and conodont *Icriodus corniger corniger* Witt. were found. The thickness is 30 m.

(25) Gray massive limestones with fewer crinoids than in units 23 and 24. Conodonts *Icriodus corniger corniger* Witt. were identified. The thickness is 14 m.

The total thickness of the *L. zmeinogorskiana* regional zone is 165 m.

The Leptagonia reticulata Regional Zone

(26) Dark-gray, dense limestones, massive, large-bedded, crinoidal, with silicified undefinable brachiopods on the weathered surface. The fauna is represented by abundant bryozoans *Amurodictya* sp., *Mongoloclema* sp. nov., *Quadrisemicoscinium planiforme* (Kop.), *Q. semirotunda* (Hall), *Medisemicoscinium alticlavatum* (Kop.), *Leptotrypella superba* Kop., and *Hemitrypa mongolica* Nekh. The following conodonts were found at the base of the unit: *Polygnathus costatus partitus* Klapp., Ziegl. et Mash., *P.* cf. *zieglerianus* Wedd., *P.* cf. *robusticostatus* Bisch. et Ziegl., *Icriodus* sp., *Belodella* sp., and *Icriodus* ex gr. *trojani* Johns. et Klapp. The thickness is 105 m.

(27) Gray, silicified limestones with tentaculitids. The thickness is 15 m.

(28) Reddish sandy limestones and siltstones alternating with dense, dark-gray limestones containing tetracorals *Cyathophyllum* (*Orthocyatus*) sp. nov. and *Heliophyllum* sp. nov. and bryozoans *Eostenopora cyphotrypoides* Kop., *Neotrematopora coreacea* Kop., *Semicoscinium alticlavatum* Kop., *S. striatum* Nekh., *Hemitrypa mongolica* Nekh., and *Pennіretepora tenuistriata* Hall. The thickness is 55 m.

The total thickness of the *L. reticulata* regional zone is 170 m.

The total thickness of the Tsagan-Khalgin Horizon is 335 m.

Eight hundred meters east of the described section, rocks in the upper half of unit 28 yield tetracorals *Amplexocarinia sibirica* (Zhelt.), *Syringaxon* sp., and *Ridderia dubatolovi* Spass. and brachiopods *Leptagonia reticulata* Grats., *Leptostrophiella alta* Aleks., *Atrypa* (*Planatrypa*) *depressa* (Sob.), and *Spinatrypa* (*Isospinatrypa*) *aspera* (Schloth.).

GOBI-ALTAI FORMATION

(29) A unit of bright green ashstone, conformably overlying member 28. The thickness is 50 m.

(30) Dark-green, finely clastic tuffbreccia, with argillite lenses in the upper part. The thickness is 15 m.

(31) Dark-gray argillites and siltstones, platy and flaggy, in places shaly. Calcareous argillites with thin interlayers of sandy limestones containing auloporids, undefinable brachiopods, and trilobites occur in the lower part of the unit. Beds saturated with tentaculitids *Styliolina fissurella* (Hall) are numerous. Conodonts *Icriodus regularicrescens* Bult. and *Polygnathus* sp. nov. were found in the lower part of the member. The thickness is 150 m.

Two kilometers east of the described section, basal layers of the member corresponding to unit 31 yield brachiopods *Vallomyonia danzani* Aleks. and conodonts *Polygnathus ensensis* Ziegl. et Klapp., *P.* cf. *angusticostatus* Witt., *Icriodus obliquimarginatus* Bisch. et Ziegl., and *Dvorakia* sp.

Rocks described in members 29–31 compose the basal part of a thick volcanogenic sequence referred to the Givetian and Frasnian.

Before proceeding to a stratigraphic analysis of the data presented above, we should note that the sampling for conodonts was performed thoroughly, at 5-m intervals, except in the terrigenous members, which were sampled at larger intervals. Among the 167 processed samples, conodonts were found in 44 samples. The calculation of specimens of different conodont genera in the studied section showed the icriodids to constitute 93% of the total composition. Representatives of other genera, namely, *Polygnathus, Belodella*, and *Pandorinellina*, are scarce. This is a typical icrioid conodont biofacies characteristic of shallow-water shelf deposits.

Correlation of the icrioid assemblages with the standard conodont zonal scale presents difficulties because the latter is based on the succession of *Polygnathus* species. Nevertheless, single findings of *Polygnathus* forms, including index species, made it possible to distinguish the *inversus* standard zone in the middle part of the Chulun Horizon as well as the *patulus* and *partitus* zones in the Tsagan-Khalgin sediments. The identification of conodont zones in the studied section is only approximate because of the scarce occurrence of index forms that may not exactly correspond to their evolutionary appearance. However, the occurrence of conodonts in the type section of the Chulun and Tsagan-Khalgin horizons permits intervals comparable to horizons and smaller stratigraphic subdivisions of the Salair to be distinguished, with regard for the data on the macrofauna and geological events of the basin.

The lower boundary of the Emsian was placed at the base of the Chulun Horizon (Alekseeva, 1990, 1993). This inference was based on (1) the correspondence of the studied level to the beginning of a large transgression and (2) the occurrence of tabulates and brachiopods known in the Emsian or appearing in other regions from that time in the lower part of the horizon. Moreover, tetracorals *E. aggregatum aggregatum* and "*M.*" *difficilis*, also characteristic of the Emsian sediments, were identified there. The former species was recorded in the Emsian of the Mountainous Altai and Australia, whereas the latter was recorded in the Emsian-Eifelian from the Belov to Mamontov horizons in the Salair.

Members 1–3 stand out in terms of their tetracoral compositions, which are characterized by accumulations of *M. latum* massive colonies and abundant branching colonies of *E. aggregatum aggregatum*. The same interval is also distinguished by a find of conodonts *I. altaicus* in the basal layers of the overlying member. These conodonts occur in the Mountainous Altai in the *gronbergi* Zone. Therefore, the studied interval of the Chulun Horizon (members 1–3) may correspond to the lower Emsian *dehiscens* Zone and may be correlated with the Salair Horizon of the Salair (refer to table).

Upward in the sequence, the interval of members 4–12 is also distinguished by a tetracoral composition. Here, the branching colonies of *Embolophyllum* with thin cylindrical corallites vanish, whereas *Tabulophyllum* and *Heliophyllum* appear. Tetracorals are more diverse here than in the basal layers. Forms of wide stratigraphic range with prevailing Altai-Sayan faunal elements were recorded. However, tetracorals did not form an assemblage characteristic of certain horizons from the Salair. The studied part of the section may be referred to the *gronbergi* Zone according to the presence of conodonts *I. altaicus* in member 5 and the

occurrence of the *inversus* zonal index species at the base of the overlying sequence. Therefore, this interval of the Chulun Horizon may be correlated with the Belov Horizon of the Salair, which corresponds, according to the latest data, to the *gronbergi* Zone (Elkin *et al.*, 1987).

The rest of the section of the Chulun Horizon is mainly characterized by brachiopods. The assemblage is similar to that of the lower Shandyn layers (dark-gray bedded limestone facies), according to the generic and some species compositions as well as the presence of abundant *Leviconchidiella*. Rocks of the Chulun Horizon are overlain by the Tsagan-Khalgin basal conglomerates, above which the *P. costatus patulus* index form was found. The available data provide grounds for correlating the upper part of the Chulun Horizon with the lower Shandyn layers of the Salair. The latter correspond to the *inversus* conodont zone and the lower part of the *serotinus* conodont zone (Elkin *et al.*, 1987).

The lower boundary of the Tsagan-Khalgin Horizon was established at the base of the basal conglomerate bed corresponding to the beginning of the largest Devonian transgression in Mongolia. We can conventionally suggest that this level is similar to or coincides with the base of the late Emsian *patulus* Zone. The sediments of members 20–22 probably refer to this zone. Conodonts *P. costatus patulus* were found at the base of the interval, whereas brachiopods characteristic of the *Zdimir-M. uralensis* Zone are abundant in its upper part. Both index forms of this brachiopod zone were widespread in Mongolia in the age equivalents of member 22. The studied sediments yield abundant *M. uralensis*, *L. zmeinogorskiana*, *Mes. birmanica*, *X. umbraculum*, *E. salairica*, and *U. frequens* as well as numerous rhynchonellids that are similar in shell morphology to species from the discussed zone but referred to other genera and species as a result of the study of their inner structure (Erlanger, 1992). The brachiopod composition of the interval is similar to that of the middle Shandyn layers; therefore, the sediments of members 19–22 are correlated with the latter.

According to their stratigraphic position, members 23–26 are correlated with the upper Shandyn layers. In these units, the macrofaunal remains, excluding crinoids, are rare and uniform. They are not comparable in terms of their composition. In Mongolia, the upper part of this interval contains conodonts *P. costatus partitus*, the index form of the lower zone of the Eifelian. The stratigraphic range of the upper Shandyn layers of the Salair, barren of conodonts, was determined by intermediate correlations; they presumably embrace the *patulus* and *partitus* zones. However, the newly available material enables us to suggest that the *patulus* Zone cannot be incorporated into the upper Shandyn layers. In the Mongolian section, *P. costatus patulus* was found in the lower part of the Tsagan-Khalgin Horizon, and the top boundary of the corresponding zone can be placed within member 22 or at its top. This

Correlation of the Emsian and Eifelian stratigraphic schemes of Mongolia and Salair

Stage	Mongolia (Alekseeva, 1990)				Salair (Elkin *et al.*, 1987)	
	horizon	formation	regional zone	member	horizon, layers	
Givetian		Gobi-Altai Formation (lower part)		29–31	Akarach	
Eifelian	Tsagan-Khalgin		*reticulata*	27–28	Mamontov	
				26	Shandyn	upper Shandyn layers
			zmeinogorskiana	23–25		
Emsian				19–22		middle Shandyn layers
	Chulun		*galinae-nalivkini*	13–18		lower Shandyn layers
				8–12	Belov	
			tsakhiri-nicus	4–7		
				1–3	Salair	

conclusion is based on the similarity of brachiopod assemblages from member 22 and from the *Zdimir-M. uralensis* brachiopod zone. The latter, as is known, yields index species of contiguous conodont zones *patulus* and *partitus* (Khalymbadzha *et al.*, 1985; Eikhgorn *et al.*, 1983; and others). In the studied section we place the upper boundary of the *patulus* Zone at the top of member 22, taking into account its relatively small thickness and the abrupt change in lithology in the overlying sequence.

The level under consideration, corresponding to the Emsian–Eifelian boundary, is difficult to identify not only in Mongolia and the Altai-Sayan region, but in other areas as well. Thus, many investigators do not consider the lower boundary of the Middle Devonian adopted by the International Devonian Subcommission at the base of the *partitus* conodont zone to match the requirements for an international standard (Bublichenko, 1983; Elkin *et al.*, 1985; Rzhonsnitskaya, 1983; and others).

The upper part of the Tsagan-Khalgin Horizon (members 27 to 28) is correlated with the Mamontov sequence according to the occurrence of tetracorals *A. sibirica* and *R. dubatolovi*. These species are distributed in the Bobkov layers of the Rudnyi Altai, which are correlated with the upper half of the Mamontov Horizon. According to the earlier notion, the Eifelian upper boundary was placed at the base of the volcanogenic Gobi-Altai Formation (Alekseeva, 1993). But in that case, the thickness of sediments corresponding to all other Eifelian conodont zones constitutes only 70 m. It is not inconceivable that the Tsakhirin-Khuduk section is stratigraphically incomplete due to a tectonic deformation at the base of the Gobi-Altai sequence. The basal sediments of member 31 yield conodonts, which are referred to the lower part of the Givetian. They occur together with brachiopods *V. danzani*, which were encountered in the Mandal-Gobi Massif with the following conodonts, also characteristic of the lower part of the Givetian: *Polygnathus linguiformis linguiformis* Hinde, *P. linguiformis alveolus* Wedd., *P.* cf. *trigonicus* Bisch. et Ziegl., *Icriodus obliquimarginatus* Bisch. et Ziegl., *I. brevis* Stauf., and *Ozarkodina semialternans* (Wirth). Therefore, we currently do not have enough data to solve the question of the position of the Eifelian upper boundary. This problem needs further investigation in the Somon Shine-Dzhinst area, as no equivalent sections have been discovered in other regions.

CONCLUSION

(1) The Emsian lower boundary is placed at the base of the Chulun Horizon, whereas the upper one is placed at the top of *Zdimir-M. uralensis* brachiopod beds. The Eifelian upper boundary probably coincides or approximates the level of the *V. danzani* beds, or the base of the Gobi-Altai Formation.

(2) The volumes of regional zones, which can be traced throughout the distribution area of terrigenous-carbonate sediments, were elucidated. For instance, the *U. tsakhirinicus* regional zone corresponds to the *dehiscens* and the lower part of the *gronbergi* conodont zone; the *S. (S.) galinae–C. nalivkini* regional zone correlates with most of the *gronbergi–inversus* and the lower part of the *serotinus* conodont zone; the *L. zmeinogorskiana* regional zone equals the *patulus* and the lower part of the *partitus* conodont zone.

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Reorganization of Marine Biota during the Mid-Early Permian Epoch

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Abstract—The evolution history of fusulinids, brachiopods, ammonoids, and conodonts shows that while advancing across the Sakmarian–Artinskian boundary, all these faunal groups suffered a crisis that was followed by a peak of their speciation during the Kungurian (Bolorian) Stage. The regenerated Artinskian–Kungurian biota differs from that of the Asselian–Sakmarian time by a higher taxonomic diversity, and by the first appearance of the Late Permian elements. The critical moment coincides with a period of global regression, while the active development of rejuvenated biota and its expansion correspond to the transgression stage. The climatic changes were evidently another important factor, which affected the evolutionary trend of the studied fossils.

Key words: *fusulinids, brachiopods, ammonoids, conodonts, regression, transgression, Sakmarian Stage, Artinskian Stage, Yakhtashian Stage*

Recently, one of us (Leven, 1993a) recognized that the stages in fusulinid evolution are dependent on the transgressive-regressive cycles. Two first-order cycles were distinguished in the Permian history of the Tethys, and the Yakhtashian was determined as separating the regressive and transgressive regimes. Analyzing the cyclicity outside the Tethyan region, Leven (1994) concluded that it is of a global rank, and should consequently influence not only the fusulinid evolution, but also other groups of fossils. The present study is devoted specifically to this issue. In addition to fusulinids, we analyzed pertinent stages in the evolution of brachiopods, ammonoids, and conodonts using data, which characterize not only Tethys, but other basins as well—specifically, those of East European and North American provinces.

Inasmuch as the tracing of geological events in vast areas is highly dependent on the precision of stratigraphic correlation between sections which recorded these events, we must first consider data on the critical point between the regressive and transgressive phases distinguishable in the above regions (Fig. 1). In the Tethyan region, it corresponds to the Yakhtashian Stage. The Sakmarian and tightly related Asselian deposits always occur here below the unconformity. The rocks above it are most often of the Bolorian age. Less frequently, they correspond to the Kubergandian Stage, and sometimes the basal beds of the transgressive sequence can be considered (with minor reservations) as late Yakhtashian in age.

The Yakhtashian Stage of the Tethyan scale is usually thought to be correlative with the Artinskian Stage. This is based on finds of the late Artinskian ammonoids yielded by the uppermost Yakhtashian horizons in the stratotype section of the latter stage (Leven *et al.*, 1992), and on the Artinskian conodont and ammonoids occurring together with the Yakhtashian fusulinids (*Pamirina, Darvasites*, and *Chalaroschwagerina vulgaris*) in some sections of southern China (Zhang Lixin *et al.*, 1988; Zhou Zuren, 1989). However in the Puan sections, the early Artinskian ammonoids associate with fusulinids (*Sphaeroschwagerina glomerosa* and *Rugosofusulina*), which are more archaic as compared to the Yakhtashian forms. Consequently, the lower boundaries of the Yakhtashian and Artinskian stages can be discordant, the former corresponding to a higher level. In this case, the Yakhtashian Stage would approximately correspond to the upper Artinskian substage of the Uralian scale. Then, in terms of standard scale units, the regression in the Tethyan regions can be referred to the middle of the Artinskian Stage, and the initiation of the late Yakhtashian transgression can be dated as the terminal event of this stage. The Uralian sections indirectly support this conclusion as well, because basal fine-clastic rocks of the Artinskian Stage are replaced here upward by coarse pebble-boulder deposits of its middle unit (the Irgina and Sarga horizons), which indicates a general uprising and basin shoaling. This regressive trend is swapped at a higher level by a transgressive clay-carbonate sequence of the Sarana Horizon, the terminal one in the Artinskian Stage (Chuvashov *et al.*, 1990).

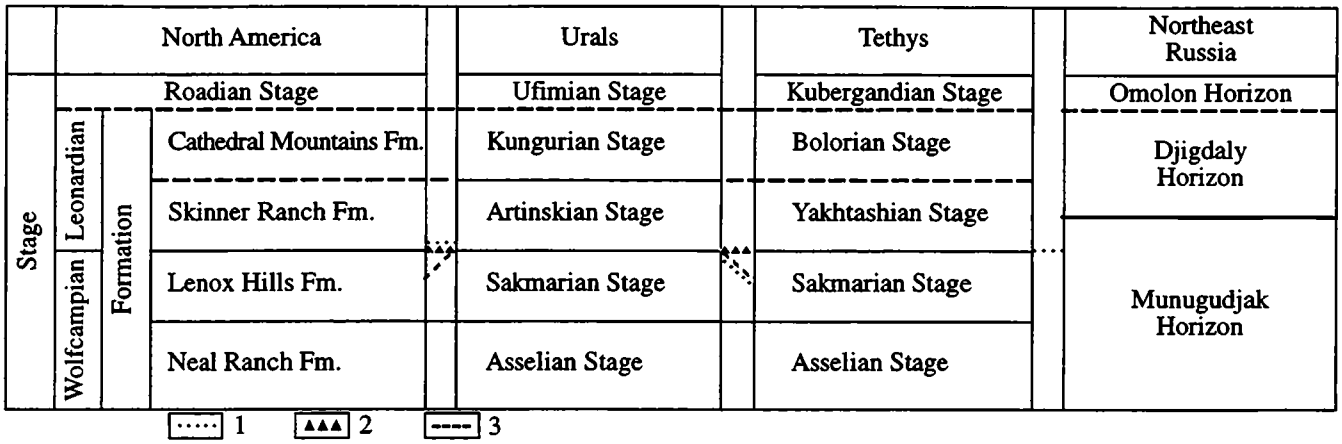

Fig. 1. Correlation between different stratigraphic scales of the Lower Permian (1, ammonoids; 2, conodonts; 3, fusulinids).

In sections of northeastern Asia, the critical level separating the cycles corresponds to the boundary between the Munugudjak and Djigdaly horizons, correlative with the middle Artinskian substage. This is evident from the fact that an upper part of the Djigdaly Horizon yielded early Artinskian (*Neoshumardites triceps* Ruzh.) and Kungurian (*Neouddenites, Daraelites*, and *Epijuresanites*) fossils (Ganelin, 1984; Ganelin *et al.*, 1990). The transgressive sequence of the North Pechora zone begins with the Gusinaya and Bel'kovskaya formations of the late Artinskian age (Bogomazov *et al.*, 1984). In Spitsbergen, a larger part of the Artinskian Stage is composed of gypsum-bearing deposits. The Kapp Starostin Formation transgressively overlying this deposits yielded conodonts of the transitional Artinskian–Kungurian age (Szaniawski and Malkowski, 1979). In the Canadian Arctic, margins of the Sverdrup basin suffered uprising movements during the second half of the Artinskian Stage (the Melvillian disturbance); the next stage of subsidence was at the very end of the Artinskian, or during the Kungurian Stage (Beauchamp, 1993).

In the Mid-Continent basin of North America, both events under consideration left behind a remarkable surface of unconformity, separating the Wolfcampian and Leonardian series. The first of them is subdivided into the Neal Ranch and Lenox Hills formations of the Glass Mountains. The uppermost horizons of the latter yielded the lower Artinskian assemblage of conodonts. In basal beds of the Leonardian Series (the Skinner Ranch Formation above the unconformity), this group of fossils is of the late Artinskian age (Movshovich *et al.*, 1979; Ritter, 1986). Data on conodonts are somewhat discordant relative to the data on ammonoids from the Lenox Hills Formation, according to which, the latter is of the Sakmarian age.

Sakmarian ammonoids were also discovered in the Hueco Formation of the same mountains (Williams, 1963). The overlying Cerro Alto Formation corresponds to a greater part of the Lenox Hills Formation in the Glass Mountains (Williams, 1963; Ross, 1963). This favors the viewpoint of Furnish (1973), who suggested the Sakmarian age of the latter. Thus, the data above need to be better correlated.

Summarizing all the information, we may state that in the principal basins of the Permian marine sedimentation, the transition from the regressive to transgressive stage was nearly synchronous worldwide, and corresponded to the second half of the Artinskian Stage and the initial interval of the Kungurian Stage. The regression maximum is confined to the middle Artinskian interval, or, when estimated more precisely, to the first half of this stage. Despite some discrepancy of dating, it is clear that the geological events under discussion were of a global rank, and therefore, should influence marine biota evolution everywhere. The faunal groups under analysis, i.e., the benthic fusulinids and brachiopods coupled with pelagic ammonoids and conodonts, were the most abundant in the Permian seas, and their evolutionary trends during the mid-Early Permian should conventionally reflect the evolution of marine biota in gross.

FUSULINIDS

This well studied group of very abundant and widespread marine fossils of the Late Paleozoic is a perfect object to study the character, scale, and dynamics of biotic changes during the selected interval of geological history. However, it is difficult to analyze fusulinid evolution as a whole, because they show individual evolutionary patterns in different biochores and a provincial character of distribution. Therefore, we consider this problem below, using the principal Permian biochores as examples.

Tethys. Fusulinids in this basin were very abundant and taxonomically diverse. As it was earlier demonstrated by Leven (1993a, 1993b, 1994), their communities suffered significant changes passing across the

Sakmarian–Yakhtashian boundary. These changes were as follows:

(1) The species composition was completely rejuvenated.

(2) Alternation of dominants was typical at both the generic and specific levels. Characteristic Asselian and Sakmarian representatives, such as genera *Sphaeroschwagerina, Pseudoschwagerina, Paraschwagerina, Zellia, Rugosofusulina, Dutkevitchia,* and *Quasifusulina* became extinct, or considerably limited in abundance. Dominant taxa were genera *Chalaroschwagerina* and *Praeskinnerella*, which appeared for the first time, *Darvasites*, which existed earlier, and new species of pseudofusulines, such as *Pseudofusulina kraffti, P. fusiformis*, and some other forms.

(3) Several newborn genera (*Pamirina, Mesoschubertella*, and *Toriyamaia*) were taxa characterizing only the second half of the Early Permian.

(4) Some new genera, which appeared during the Yakhtashian and Bolorian stages, became ancestors of higher taxa prosperous later, during the Late Permian. These were: *Misellina*, the first genus of the order Neoschwagerinida; *Yangchienia*, representing the monotaxonic family Yangchienidae, and such genera as *Nankinella* and *Staffella*, both ancestral in the Nankinellidae and Staffellidae families.

(5) The most prominent changes in Early Permian fusulinid assemblages were recorded in the southern peri-Gondwanan part of the Tethys, where the endemic Sakmarian complex of relatively cryophilic species was replaced by typical thermophilic forms (Leven, 1993b).

The dynamics of Tethyan fusulinid evolution is illustrated by a series of histograms (Fig. 2) showing variations in the amount of genera during the Early Permian. As one can see in the first histogram, the Sakmarian–Yakhtashian boundary marks the peak extinction of older genera associated with the maximum of newborn taxa, the latter promoting general growth of their abundance.

Other histograms (Fig. 2, parts *2* and *3*) well illustrate the rejuvenation degree of fusulinid assemblages. The first shows the percentage of newborn genera relative to their total number during a particular time interval. It is clear that the rejuvenation peak (59%) was during the Yakhtashian Stage. For comparison, respective values are 48% for the Carboniferous–Permian, and 40% for the Early–Middle Permian boundary intervals. Ratios of newborn genera to their total amount at the preceding stage are even more indicative: 75% for the Yakhtashian Stage, and only 60% for the Asselian and 50% for the Kubergandian stages. The last histogram (Fig. 2, part *3*) illustrates percentage ratios between the genera extinct before the upper boundary of a stage and all the genera that existed at this stage. Here, one can easily see that fusulinids suffered two significant crises during the Early Permian—first, at the end of the Sakmarian Stage (40% of genera), and next, at the end of the Bolorian Stage (45% of genera).

Boreal regions. Connections between the Boreal and Tethyan provinces were either interrupted, or irregular since the Sakmarian time. These discriminated fusulinid assemblages inhabited respective basins. As compared to the Tethyan communities, the Sakmarian and Artinskian fusulinids from the Arctic regions of Canada, Spitsbergen, the Urals and the Timan-Pechora area are distinctly less diverse. Representatives of the genus *Pseudofusulina* are obviously dominant among them, and associate with few other genera of low specific diversity.

In East European basins, the post-Sakmarian regression did not result in a complete drying of the region, but it caused a more intense supply of the basin with coarser sedimentary material (especially during the middle Artinskian time) derived from the Uralian uplifts. Accordingly, biotic changes during the period under discussion were evident here, but not as prominent as in the Tethys. For instance, Chuvashov (1984) reported that fusulinids from the Burtsevka Horizon of the Artinskian Stage resemble those from the Sterlitamak Horizon of the late Sakmarian age. Using these data, he suggested to discriminate an independent Yuresan' regiostage comprising the two units mentioned above and the Irgina Horizon overlying them. Such a new stage seems to be very conventional, because the Irgina time marks an important moment in fusulinid evolution, when the first occurrence level of their elongated forms with cuniculi (*Praeparafusulina lutugini*) were soon replaced up-section by parafusulines. It should be noted that in the Tethyan

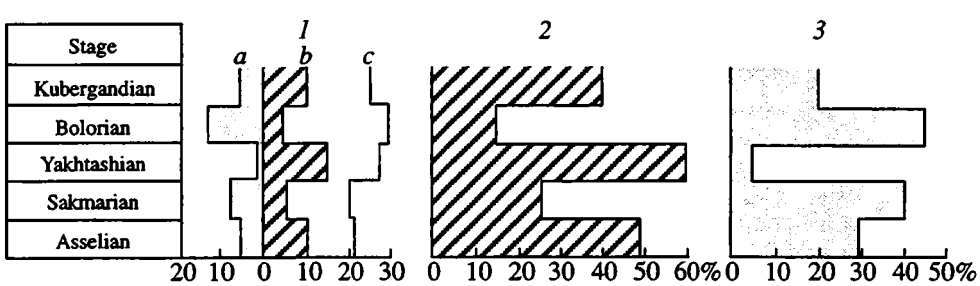

Fig. 2. Fusulinids: histogram demonstrating secular variations (*1*) in distribution of extinct (*a*), new (*b*), and all (*c*) fusulinid genera; percentage of new (*2*) and extinct (*3*) genera relative to their total amount at different stages.

regions, the first cuniculi in fusulinid forms appeared at the end of the Yakhtashian–Bolorian interval, and, according to Bogoslovskaya (1988), the Irgina assemblage of fusulinids is of the late Artinskian age, though in standard stratigraphic schemes of the Urals the Irgina Horizon it is still assigned to the early Artinskian substage.

In other Boreal basins, which were inhabited by Early Permian fusulinids (from Alaska on the west to Spitsbergen on the east), the middle Artinskian events resulted in the complete extinction of this faunal group. There are many reasons to believe that this was a consequence of regression, which left respective marks in many sections of Canada and Spitsbergen. Fusulinids of the Asselian, Sakmarian, and Artinskian age (up to the Irgina and possibly Sarga horizons) occur mostly in the carbonate Belcher Channel Formation of the Sverdrup basin. The basal third of the overlying Assistance Formation (and its analogs) contains the Roadian (=Kubergandian, or Ufimian) fusulinids, as well as conodonts. According to Kozur (1977), the latter group corresponds to the Bolorian conodonts of the southeastern Pamirs. Some researches (Nassichuk, 1975; Nassichuk and Wilde, 1977) noted a hiatus below the base of the Assistance Formation, and Chuvashov (1984) considered this hiatus as comprising the entire Artinskian Stage, the latter corresponding, in his opinion, to a stratigraphic range from the Sarga Horizon to the upper boundary of the Kungurian Stage. If we take into account conodonts, which are Bolorian (=Kungurian) in age and occur at a higher level than the base of the Assistance Formation (Kozur, 1977), then the initial level of the hiatus should be assigned to the second half of the Artinskian, or possibly, to the very beginning of the Kungurian time. A similar conclusion is inferable from data on Spitsbergen, where the Wordiekammen Limestone contains the Sakmarian (not younger) fusulinids (Ross, 1963). The gypsum-bearing deposits above this unit seem to correspond to the regression stage. The overlying basal part of the Kapp Starostin Formation yielded conodonts of the transitional Artinskian–Kungurian age (Kozur, 1978), marking the initial transgression. Paleogeographic, paleoclimatic, and paleoecologic transformations, i.e., all the results of the Artinskian regression in the Boreal province, were significant to such an extent that fusulinids were unable to survive them, and all over the province, the Upper Permian deposits are barren of these fossils.

North American regions. The Permian type sections of the Mid-Continent basin show that most significant changes in fusulinid assemblages were at the Wolfcampian–Leonardian boundary, which is often confined to a surface of unconformity. Here, similar to the Tethyan regions, fusulinid assemblages of the Asselian and Sakmarian (possibly Sakmarian–early Artinskian) stages, respectively correlative to the Neal Ranch and Lenox Hills formations, display many common features. Both assemblages include loosely spiraling forms, such as *Pseudoschwagerina* and *Paraschwagerina*. These forms disappear above the unconformity in the Skinner Ranch Formation, where dominant species correspond to highly developed representatives of the genera *Pseudofusulina, Skinnerella*, and *Parafusulina*, most prosperous during the Late Permian. The difference between the assemblages is very sharp, suggesting a significant hiatus in sedimentation separating the accumulation periods of respective host deposits. This was first noted by Ross (1963), who concluded that *Chalaroschwagerina nelsoni* and *Pseudoschwagerina gerontica* beds, distinguishable in an upper part of the Alacran Mountain Formation (the Hueco region), are missing in sections of the Glass Mountains.

BRACHIOPODS

Like fusulinids, brachiopods are benthic organisms with a highly provincial distribution. Therefore, their evolution could also be traced in separate large biochores.

Tethys. The most distinct critical point in the Tethyan brachiopod evolution is related to the Yakhtashian regression. It was of particular significance in the peri-Gondwana regions (Iran, Arabian Peninsula, southern Afghanistan, southern Pamirs, Karakorum, Himalayas, northern India, southern Tibet, and Australia), where the biotic impact during regression and subsequent transgression was significantly magnified by climatic factors. Here, the late Asselian (?)–Sakmarian communities of brachiopods generally consisted of cryophilic forms of the so-called Gondwanan type. They included approximately 20 genera, among which *Taeniothaerus, Punctocyrtella, Notospirifer*, and *Triginitreta* were the most indicative (Grunt and Dmitriev, 1973; Grunt, 1993; Archbold and Gaetani, 1993). In many of these regions, the Yakhtashian interval corresponds to a hiatus in sedimentation, and the features of respective brachiopod assemblages are unclear.

The Bolorian brachiopods are abundant in the southeastern Pamirs (Grunt and Dmitriev, 1973) and Khao Phric locality of southern Thailand (Grant, 1976). They are very different in taxonomic composition from the Sakmarian community. Only one genus is inherited from the latter, while the species list includes absolutely new forms. The much more diverse Bolorian community comprises over 70 genera, including 30 endemic taxa such as *Incisicus, Transennatia, Bibatiola, Celebetes, Haydenella, Posicomta, Cryptospirifer, Spirigerella, Boloria, Overtoniina, Uncinella*, and others. Many of them became widespread in the Late Permian basins of the Tethys. Other genera were of a wider stratigraphic and geographic distribution. It is remarkable that some genera of the latter type also occur below the Bolorian deposits in the northern Tethyan regions, as well as in the southern Urals, which was a part of the tropical climatic zone during the Asselian–Sakmarian time. This suggests that reorganizations in the peri-Gondwanan brachiopod assemblages resulted not only from ecological changes during the Yakhtash-

ian regression, but also from the migration of their thermophilic forms into the southern Tethyan regions. The latter was possible because of the climatic warming of the Yakhtashian time.

The above analysis of fusulinid assemblages implies the same conclusion. In general, the Bolorian brachiopods from the southern Tethys do not display evolutionary links with older Sakmarian forms, thus representing an absolutely new assemblage consisting of endemics and migrants from warm-water basins. The initial formation of the assemblage is likely rooted in the Yakhtashian Stage, when most respective innovations were not a consequence of evolutionary transformations, but rather, a result of the arrival of thermophilic migrants superseding cryophilic forms.

In the northern Tethyan regions, i.e., in equatorial basins of the Early Permian time, climatic factors were insignificant, and the changes in brachiopod assemblages were of a different character. According to our data on southwestern Darvaz, the late Asselian–early Sakmarian brachiopods represented, as in the southern Urals, significantly diverse reef assemblages there. Many of their genera, such as *Enteletes, Chaoiella, Tubaria, Keyserlingina, Echinoconchus, Urushtenia, Trautsholdia, Brachythyrina, Choristites, Crurithyris, Eliva, Wellerella, Stenoscisma*, and others, are similar to Uralian taxa. Nevertheless, they usually include endemic species.

In Darvaz, the upper Sakmarian and lower Yakhtashian deposits belong to terrigenous facies nearly always barren of brachiopods. These fossils appear higher in the section, at the level of the Safetdaron Formation corresponding mostly to the Bolorian Stage. In general, brachiopods from this formation are characteristic of the reef assemblages, and resemble late Asselian–Sakmarian forms. At the same time, this is the first occurrence level of many genera and species recorded in the Bolorian deposits of the southern Tethys, for instance, in the Khao Phric locality. Such features may point to a low climatic gradient between the peri-Gondwanan and equatorial regions of the Tethys.

East European basins. Despite a somewhat impoverished taxonomic composition, the Sakmarian brachiopods from these basins are closely related to the Asselian forms. In both groups, many taxa are inherited from the Middle–Late Carboniferous biota. Their characteristic representatives are the following genera: *Enteletes, Meekella, Kozlowskia, Chaoiella, Reticulatia, Tubaria, Retaria, Marginifera, Keyserlingina, Echinoconchus, Urushtenia, Callispirina, Rugivestis, Juresania, Trautsholdia, Brachythyrina, Choristites, Purdonella, Crurithyris, Eliva, Tiramnia, Federicksia, Punctospirifer, Wellerella, Stenoscisma*, and *Leiorhynchus*. These genera are typical of thermophilic communities, and in addition to bryozoan and algae, they contributed much to the formation of reef structures such as the Shikhan and Sterlitamak reefs. Many of the above genera became extinct before the end of the Sakmarian Stage. During the Asselian and Sakmarian stages, the latter in particular, these genera were associated with rare representatives of taxa, which later became important constituents of brachiopod assemblages. To exemplify such taxa, we may list genera *Chonetina, Dyoros, Arctochonetes, Kochiproductus, Achunoproductus, Jakovlevia, Anidanthus, Muirwoodia, Megousia, Waagenoconcha, Neospirifer*, and *Spiriferella*.

Though significantly impoverished, the early Artinskian (Burtsevka Horizon) brachiopod assemblages are still similar to their Sakmarian counterparts. Their rejuvenation was initiated at the late Artinskian level, and resulted in a much higher taxonomic diversity. For instance, the Sarga Horizon yielded 140 brachiopod species of 55 genera. This faunal group is enriched in chonetids (*Quadrochonetes, Dyoros, Chonetina, Tornquistia, Svalbardia, Arctochonetes*, and *Neochonetes*), spiriferids (*Cyrtella, Spiriferella, Neospirifer*, and *Torynifer*), and productids (*Megousia, Thuleproductus, Jakovlevia, Uraloproductus, Horridonia, Sowerbina, Anemonaria, Avonia*, and *Paucinispinifera*). They associate with widespread rhynchonellids (*Rhynchopora*) and terebratulids (*Dielasma* and *Beecheria*). In contrast to pre-Artinskian brachiopods, which clearly represent thermophilic communities typical of the northern Tethys, the late Artinskian assemblages included more cryophilic forms characteristic of the West Arctic basins of the Late Permian. Respectively, forms of the Tethyan type gradually became less abundant, reflecting the progressing isolation of the East European basins from the Tethys and an expansion of the Boreal fauna.

The Kungurian regression was unfavorable for the brachiopod evolution, and their assemblages from that time are less abundant and diverse than the Artinskian taxa. The Timan–Pechora region was the only exception, and the brachiopod communities here underwent the next rejuvenation stage, evolving further under preserved marine conditions.

Taimyr–Kolyma basins. Changes in brachiopod communities represent only one aspect of general ecological transformations—those which affected these regions during the mid-Artinskian time. This level marks the boundary between two large cycles of sedimentation (Kashik *et al.*, 1990), and a critical point when fauna of one ecological type was replaced by another. At this very moment, older faunal groups of the Verkhoyansk type were replaced by extremely different fauna of the Djigdaly type. This general reorganization is evident in all biotic groups (Ganelin and Kotlyar, 1984a, 1984b).

Changes in brachiopod assemblages are the most thoroughly studied. The Asselian–early Artinskian brachiopod communities included eight families of Rugosochonetidae, Anopliidae, Overtoniidae, Linoproductidae, Rhynchoporidae, Neospiriferidae, Elithidae,

Amboco[illegible]idae, and Pugnacidae. Three of them (Anopliidae, [illegible]rtoniidae, and Pugnacidae) totally disappeared fr[illegible] the basins of northeastern Asia at the stratigraphic l[illegible]el under discussion. Seven other families, including such important groups as Meekellidae, Buxtoniidae, Horridoniidae, and Licharewiidae, appeared to replace them. Changes at the generic level were equally essential. For instance, as many as nine genera disappeared from the assemblage of 16 genera typical of the Munugudjak Horizon, among them such taxa as *Tornquistia* and *Jakutoproductus*, previously dominant among the others. New taxa of the Djigdaly time include nearly 20 genera, among which *Meekella, Lissochonetes, Waagenoconcha, Sowerbina, Megousia, Pseudosyringothyris, Tumarinia, Timaniella, Alispiriferella, Crassispirifer*, and *Tomiopsis* were the most abundant. One more feature is a higher diversity at the specific level (up to 45 species, nearly all rejuvenated).

It is important that outside the region, for instance in the East European basins, many genera and species of the Djigdaly Horizon occur in older deposits as well. Their appearance in northeastern Asia shows that the Djigdaly transgression and related lowering of climatic gradients were favorable for distant migrants. From this moment on, the biota of northeastern Asia was acquiring features common to the coeval faunas of the main Arctic basins to finally become considerably less endemic than before.

North American regions. During the Early Permian, the Mid-Continent basin was in the tropical zone, and accordingly, the transition from Sakmarian to subsequent brachiopod assemblages was more gradual here than in the southern Tethys. In addition, the post-Sakmarian regression was short and not very sharp in this area. Nevertheless, like everywhere else, the Artinskian Stage was critical for the brachiopod evolution even here. Their remains are studied most thoroughly in the Glass Mountains (Cooper and Grant, 1972). A series of new forms was discovered in this locality from the Skinner Ranch Formation, which is correlative to the Artinskian Stage of the Urals according to ammonoids, or to its upper substage according to conodonts. The first occurrence level for such genera as *Acolosia, Chonosteges, Coscinophora, Cyclacantharia, Entronaria, Cartorhium, Oncosarina, Rhamnaria, Rugaria*, and *Thamnosia*, is at the formation base. The first specimens of the *Paranorella* genus were recorded in its middle part, while the upper part yielded representatives of *Sceletonia, Torynechus, Gliptosteges, Lepidospirifer, Phrenophoria, Plectelasma, Elassonia*, and *Tschernyschewia* genera. Many of the listed genera occur only in the Skinner Ranch Formation, though some of them are also widespread in the overlying Cathedral Mountain Formation.

Despite the fact that brachiopods from the Skinner Ranch Formation are enriched in new forms absent in the older Lenox Hills Formation, both formations are closely interrelated and contain many common genera, such as *Scachinella, Derbyia, Eolyttonia, Geyerella, Kochiproductus*, and some others. Cooper and Grant considered distribution of this kind to be important enough to refer the Skinner Ranch Formation to the Wolfcampian Group. However, it seems to be more correct to detect stratigraphic levels on the basis of newly appeared taxa, especially when such changes are also typical of coexisting fusulinids and ammonoids, which are even more pronounced among the two latter groups. Upsection of the Skinner Ranch Formation, most of its older Wolfcampian elements of fauna disappear, and brachiopod assemblages from the Cathedral Mountain Formation are absolutely rejuvenated in comparison with those from the Lenox Hills Formation.

AMMONOIDS

Pelagic ammonoids are more cosmopolitan as compared to fusulinids and brachiopods. The principal features of their evolution were similar in different basins. Formed at the beginning of the Early Permian, this faunal group persistently evolved toward greater taxonomic diversity up to the end of the Bolorian Stage, when the most characteristic taxa of the Early Permian time underwent a global extinction. The general course of evolution was not very regular, as is evident from the histograms in Fig. 3. They show that the initially smooth diversification of ammonoid genera was disturbed by a speciation outbreak during the Artinskian Stage. This outbreak was preceded by an extinction event of the terminal Sakmarian time, which was the most significant for the entire Early Permian interval. Changes at the Sakmarian–Artinskian boundary simultaneously modified the abundance and diversity of ammonoids. Many genera of Carboniferous origin,

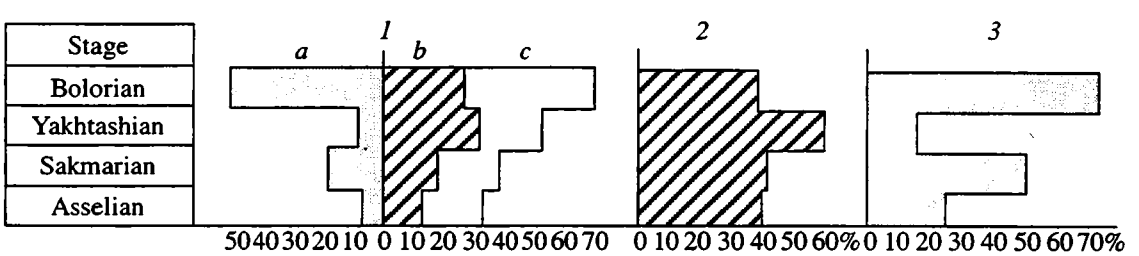

Fig. 3. Ammonoids: histograms demonstrating secular variations (*1*) in distribution of extinct (*a*), new (*b*), and all (*c*) fusulinid genera; percentage of new (*2*) and extinct (*3*) genera relative to their total amount at different stages.

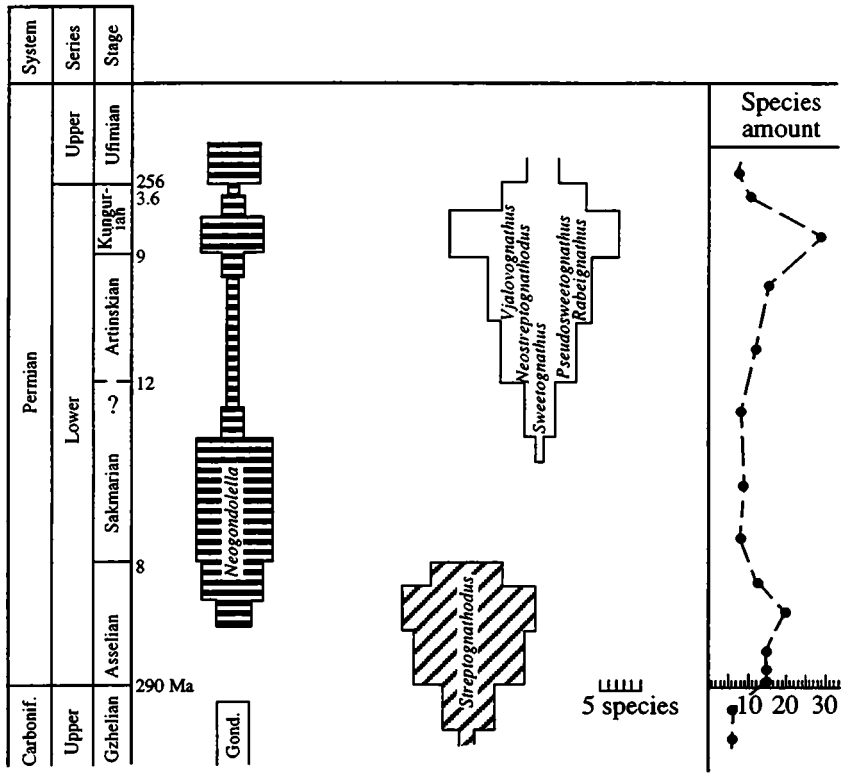

Fig. 4. Species diversity of main Early Permian conodont genera.

which were widespread during the initial interval of the Early Permian, ceased to live during the Sakmarian Stage. These were *Boesites, Metapronorites, Eoasianites, Glaphyrites, Somoholites*, and *Emilites*. Only four long-lived genera (*Neopronorites, Artinskia, Neoaganides*, and *Agathiceras*) survived this stage to become significant components of subsequent ammonoid biota. The beginning of the Artinskian Stage was the first occurrence level for *Daraelites, Eothinites, Popanoceras, Waagenina*, and *Perrinites* genera, the taxa characteristic of the entire second interval of the Early Permian. Simultaneously born genera *Aktubinskia, Artioceras*, and *Neoshumardites* were of a lesser vitality, and died at the end of the Artinskian Stage. All listed genera indicating the commencement of this stage initially had a very limited distribution and abundance like other ammonoid assemblages of that time. The only exception was the biota of the southern Urals (Ruzhentsev, 1956), which included very abundant and diverse ammonoids of the early Artinskian age (Burtsevka and Aktasta horizons). Their intense speciation was confined to the mid-Artinskian time (Bogoslovskaya, 1988). The well-known and rich Artinskian fauna of ammonoids attained its final complexion precisely at this moment. The Urals and Tethys were principal areas of its origin, diversification, and distribution.

An intense speciation and expansion of ammonoid fauna during the mid-Artinskian time continued through the Bolorian (Kungurian) Stage, when over 20 new genera appeared in ammonoid communities (Andrianov, 1985; Leonova and Dmitriev, 1989; Leven *et al.*, 1992). However, many of these genera were limited in distribution by their source regions.

The late Bolorian (late Kungurian) crisis marked the terminal moment in the history of the Early Permian ammonoids, because nearly all their genera of that period ceased to live, regardless of the time of their origin, and completely disappeared. After the mass ammonoid extinction at the Bolorian–Kubergandian boundary, paraceltites (the Permian ceratites) and goniatites with the most complex septa (Kufengoceratidae and Cyclolobidae) were the main *personages* of the Permian history of ammonoids.

Thus, the mid-Early Permian event at the Sakmarian–Artinskian boundary was the principal one in the history of ammonoid fauna, and resulted in cardinal biotic changes which influenced its further evolution. This event subdivided the history into two large intervals with similar evolutionary trends. The short initial period of each interval was the phase of biota formation (the initial phase of the Asselian Stage and the Burtsevka time of the Artinskian Stage), when ammonoid groups were small and included many older forms. These were also the origin phases for large taxa, whose rapidly evolving representatives created the basis of subsequent biotas. Such taxa included four families (Paragastrioceratidae, Metalegoceratidae, Perrinitidae, and Popanoceratidae) and five genera (*Mescalites, Juresanites, Prostacheoceras, Kargalites*, and *Tabanta-*

lites) at the beginning of the Asselian Stage, and one family (Eothititidae) with eight genera at the initial phase of the Artinskian Stage. Short periods of biota formation were followed by speciation and expansion outbreaks which more or less lasted until the deep final crisis.

It should be noted that the description of ammonoid evolution presented here is too generalized, because ammonoid finds and their complete successions in geological sections are too sporadic. The description is chiefly based on data from the Urals, where the transition from Sakmarian to Artinskian ammonoids is most evident, and traces of the Artinskian regression are insignificant (Ruzhentsev, 1951, 1952, 1956; Bogoslovskaya, 1962, 1988; Chuvashov and Bogoslovskaya, 1981). In other regions, which were more intensively affected by this regression, the difference between the Sakmarian and post-Sakmarian assemblages of ammonoids may contrast even more (Leven *et al.*, 1992; Furnish and Glenister, 1977).

CONODONTS

The Early Permian history of conodonts is distinctly divisible into three intervals (Fig. 4), when their dominant forms were successively represented by streptognathoids, neogondolelloids, and neostreptognathoids (Barskov and Reimers, 1992). The first of them corresponded to the Asselian Stage, with abundant conodonts of the genus *Streptognathodus*. The next interval corresponding to the Sakmarian Stage is marked by the predominance of the *Neogondolella* species, whose first occurrence level is in the Asselian deposits. The maximum diversity of this genus is confined to the first half of the Sakmarian Stage, when it included at least nine species, all represented by larger forms. The crisis in their evolution was at the end of the stage, and only one species, *Neogondolella bisselli*, survived this event (Ritter, 1986).

The origin of new genera *Sweetognathus* and *Neostreptognathodus*, coeval with the decline phase of *Neogondolella*, symbolized the beginning of the third period in conodont evolution. In the Urals, the first primitive forms of the genus *Sweetognathus* appeared in the uppermost Sakmarian horizons. The first conodont specimens with deep axial sulculus, such as *Neostreptognathodus obliquidentatus*, *N. pequopensis*, and *N. exculptus*, were recorded at the base of the Artinskian strata (Chernykh, 1989). Here, they are associated with the widespread species *Sweetognathus withei*. In Tethyan regions, these species occur at the base of the Yakhtashian Stage, or immediately below it (Zhou Zuren, 1988; Zhang Linxin, 1988). According to the fusulinid scale, this lower level should be assigned to the Sakmarian Stage, however, according to conodonts, it is correlative with the lowermost Artinskian beds of the Urals, and this solution seems to be more correct. In North America, as in the Urals, the most primitive *Sweetognathus* forms occur somewhat lower than *Sw. withei* and the first representatives of the genus *Neostreptognathodus*. The last forms were reported from the uppermost part of the Lenox Hills Formation bearing, as was already mentioned, the Sakmarian ammonoids.

Thus, the data above suggest that the foremost elements of the *Neostreptognathodus* assemblage signifying the new stage in conodont evolution appeared at the Sakmarian–Artinskian boundary, a little before the rejuvenation stage distinguishable in fusulinid and brachiopod communities. The initiated changes in conodont assemblages underwent intense progress during the late Artinskian–Kungurian time, and resulted in a high diversity of *Sweetognathus* and *Neostreptognathodus* species. New genera *Pseudosweetognathus, Rabeignathus, Vjalovognathus*, and *Homeoiranognathus* also appeared during this interval. Another outbreak of speciation was typical of *Neogondolella* forms at the very end of the Artinskian time and throughout the Kungurian Stage. The maximum taxonomic diversity of conodonts is confined to the Kungurian (Bolorian) Stage. At the end of this interval, representatives of the *Sweetognathus* and *Neostreptognathodus* genera were under decline, whereas *Neogondolella* forms continued their evolution to become basic elements of the Late Permian conodont communities.

CONCLUSION

Our analysis of the Early Permian evolution of four main groups of marine fauna suggests the following conclusions:

(1) Despite different habitat environments, all the groups underwent an evolutionary crisis during the mid-Early Permian time, which was followed by a rejuvenation stage, when new elements typical of the Late Permian appeared in all communities.

(2) The crisis was confined to the Sakmarian–Artinskian boundary interval. The first results of rejuvenation are notable in the initial–middle interval of the Artinskian Stage, while the speciation outbreak was at the end of this stage and during the Kungurian (Bolorian) time. Having a great deal in common, the Asselian and Sakmarian assemblages show many features inherited from their Carboniferous ancestors. The Late Permian elements in rejuvenated assemblages of different groups were of dissimilar evolutionary significance.

(3) Other groups of fauna also experienced comparable changes during the stratigraphic interval under consideration (Astaf'eva and Astaf'eva-Urbaitis, 1988; Guseva and Koshevarova, 1984). These changes were most evident in the Boreal and Notal provinces, where transformations affected the biota altogether. For instance, during the Asselian–Sakmarian time, the Sverdrup basin of the Canadian Arctic was inhabited by tropical organisms (chlorosponge and chloroforam) with dominant groups represented by algae, foraminifers (fusulinids including), calcareous sponges, colo-

nial corals, and paleoaplisines. During the Artinskian and Kungurian stages, they were replaced first by moderately thermophilic (brionoderm-extended), and then by cryophilic (brionoderm) assemblages, the latter consisting of bryozoan, brachiopods, crinoids, and bivalves (Beauchamp, 1993). The same changes were also typical of the Timan–Pechora basin. Equal in significance, opposite evolutionary patterns were recorded in the peri-Gondwana regions of the Tethys, where the cryophilic Asselian community of the eurydesmo-conulariid type with associated bryozoan and brachiopods was replaced by the thermophilic assemblages of the late Yakhtashian–Bolorian time, which consisted mostly of fusulinids, corals, brachiopods, and different algae, the latter also represented by reef-building taxa.

(4) The crisis coincided with the initial moment of a global regression, which was at its maximum during the middle Artinskian interval. The late Artinskian rejuvenation of biota seems fit Boucot's principle (Boucot, 1975), which suggests that rates of new taxa generation are increasing when their habitat areas are decreasing during regressive cycles. However, the late Artinskian–Kungurian radiation of rejuvenated assemblages definitely corresponds to the transgression cycle.

(5) In addition to the direct transgressive–regressive influence, the biota evolution was evidently under the control of related paleogeographic changes that ceased, or opened connections between basins, thus transforming the migratory ways of fauna. The examples are the late Artinskian–Kungurian migration of some bivalve taxa from the Notal into the Boreal basins (Astaf'eva and Astaf'eva-Urbaitis, 1988), and the coeval expansion of boreal brachiopods into the northern Tethys (Grunt and Dmitriev, 1973).

(6) Climatic changes were also significant factors affecting the biota evolution during the stratigraphic interval under consideration. Their influence was most evident in basins located within high latitudes, where climatic fluctuations were the most contrasting.

(7) The late Sakmarian–early Artinskian biotic crisis and subsequent rejuvenation of biota are the results of many factors, and we do not know so far which one was the main and most influential among them. Nevertheless, it is clear that the Artinskian Stage corresponds to the critical moment of the Permian geological history, implying that the biotic and abiotic events of this stage should be evaluated to a greater extent than they have been.

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Upper Pliocene–Lower Pleistocene Marine Deposits of Western Syria: Stratigraphy and Paleogeography

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Abstract—Complex characteristics of upper Pliocene–lower Pleistocene marine deposits filling the deepest part of the Al-Kebir depression near Lathakia, western Syria, are given. New biostratigraphic and paleomagnetic data allowed a detailed subdivision of the upper Pliocene–lower Pleistocene marine sequence. The position of the Pliocene–Pleistocene boundary in association with the Olduvai paleomagnetic event is outlined. The post-Pliocene marine deposits are 30–40 m thick and occur at a level of 200 m, which is considered to correspond to the Calabrian terrace. The maximum of the Pliocene transgression was confined to the early Pliocene and the beginning of the late Pliocene. At the end of the Pliocene, a gradual regression occurred. In the early Pleistocene (1.65–0.8 Ma), a shallow marine basin existed in western Syria. Palynologic data and paleotemperature estimates based on the Ca/Mg ratio in marine mollusk shells indicate a cooling at 2.4 Ma. The late Pliocene–early Pleistocene was characterized by polycyclic climatic oscillations that occurred against a background of progressive aridization and relative cooling.

Key words: *stratigraphy, paleogeography, Pliocene, palynology, foraminifers, nannoplankton, paleomagnetism*

The stratigraphy of Pliocene–Quaternary marine deposits of the eastern Mediterranean is mainly based on studies of deep-water sequences, whereas coastal marine sections are poorly studied. Coastal marine sequences suitable for biostratigraphic and paleomagnetic studies of upper Pliocene–lower Pleistocene deposits are rare in the Mediterranean except for the well-studied Italian localities. The best sequences for this purpose are located in western Syria near Lathakia, where marine Pliocene deposits fill the deepest southwestern part of the Al-Kebir depression (Fig. 1).

Studies of marine micro- and macrofossils provided the basis for the biostratigraphic subdivision of the Pliocene deposits of western Syria. Based on detailed investigations of foraminifers, the Pliocene in the coastal region of Syria was subdivided into lower and upper units (Krasheninnikov, 1971). Upper Pliocene sediments correlative with the Piacenzian contain rich assemblages of malacofauna, which were studied by Ronger, Roman, and Merklin.

In addition, much attention was given to studies of Pliocene marine and alluvial terraces in coastal areas of Syria and Lebanon (Besancon *et al.*, 1977; Copeland and Hours, 1978; Besancon, 1981a, 1981b; Sanlaville, 1981).

Despite this, the upper Pliocene–lower Pleistocene boundary interval remains poorly studied. The problem of defining the Neogene–Quaternary boundary in these sections has been insufficiently investigated. As is known, in the Vrica stratotype section in southern Italy, the latter is recommended to be drawn at the level of 1.65 Ma (Colalongo *et al.*, 1986; Nikiforova *et al.*, 1986; Rio *et al.*, 1991).

New data on the geochronology and paleogeography of the terminal stage of the Pliocene transgression and on the transition to the Pleistocene transgressive–regressive phases of the East Mediterranean basin are of a great importance. This paper deals with the materials collected from the upper Pliocene–lower Pleistocene sequences in coastal areas of Syria during brief field excursions in 1990–1992. These Russian–Syrian expeditions included E.V. Devyatkin, A.E. Dodonov, K. Khatib, and H. Nseir as participants.

GEOLOGIC STRUCTURE OF PLIOCENE–QUATERNARY DEPOSITS AND GEOMORPHOLOGY OF COASTAL AREAS OF SYRIA

Pliocene marine deposits are spread in the southwestern part of the Al-Kebir depression, which had formed by the beginning of the Neogene and was filled

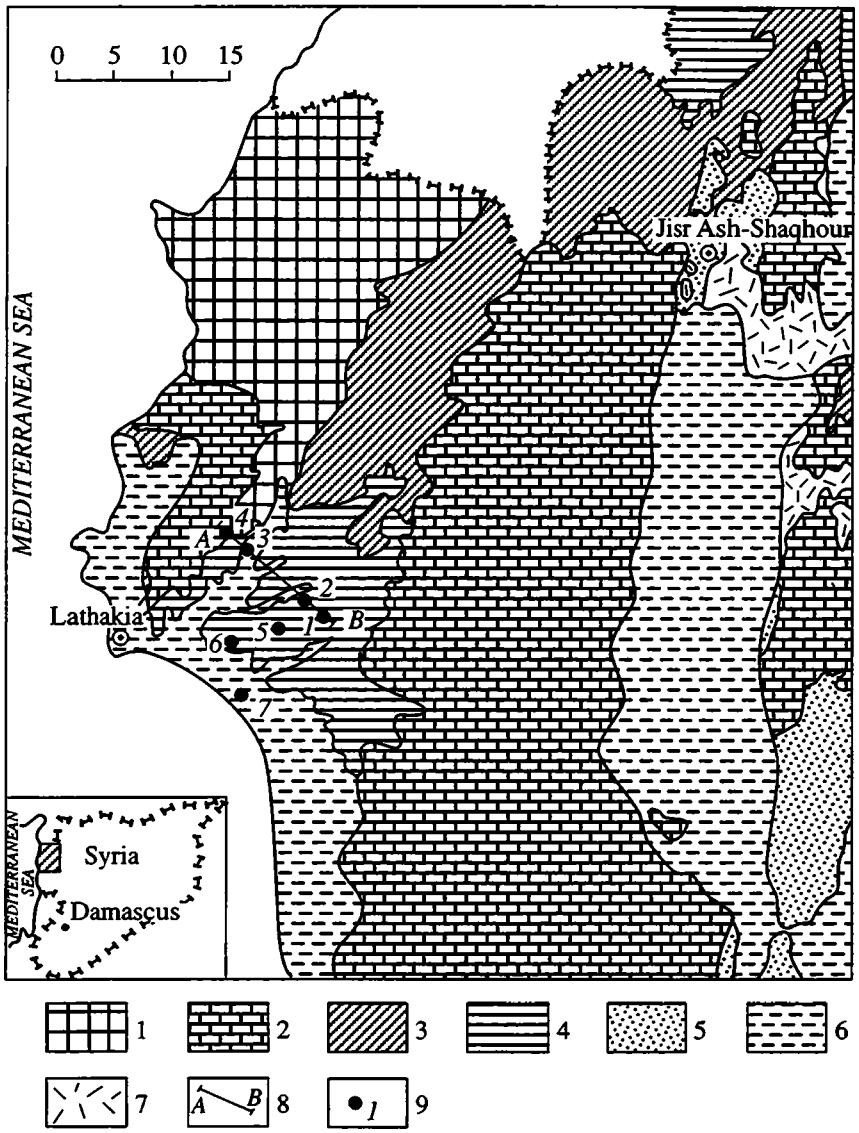

Fig. 1. Schematic geological map of the studied area in western Syria. 1, Mesozoic ophiolites; 2, Jurassic-Cretaceous–Paleogene carbonate rocks; 3, Miocene terrigenous-carbonate deposits; 4, Pliocene marine clays; 5, Pliocene continental deposits: conglomerate, sand, clay; 6, Quaternary sediments: pebblestone, sand, clay; 7, Pliocene-upper Pleistocene basalts; 8, position of geologic cross-section along line *A–B*; 9, studied geologic sequences of Pliocene–Quaternary deposits: (*1*) Msherpheh, (2) Mardido, (*3*) Jinndiriye, (*4*) Mshairfet-es-Samouk, (*5*) Al-Quatriya, (*6*) Hennadi, (*7*) Snobar.

with Miocene terrigenous-carbonate and gypsum-bearing sediments. Pliocene clays transgressively overlie Miocene layers of various ages. The boundary between the Messinian gypsum-bearing member and the Pliocene clays is uneven. The top layers of the former bear deep (up to 5–8 m) leaching hollows, and the basal layers of the latter contain pebbles and coarse clastic material. The leaching hollows implay the surface exposure of gypsum and a hiatus at the Messinian–Pliocene boundary. The Pliocene deposits are composed of gray and bluish-gray clays with varying admixtures of sandy material; sometimes they are marly with separate limestone beds. The total thickness of the Pliocene member is about 300–400 m. The most detailed studies were concentrated in the upper part of the Pliocene clayey member of the Mardido, Msherpheh, Al-Quatriya, and Jinndiriye sequences (Fig. 1). Artificial terraces on slopes made it possible to study the 100-m-thick sandy-clayey member of the Mardido and Msherpheh sections, located near the villages with corresponding names. Lithologically, this member consists of rather uniform silty clays with rare sandy beds. In the top layers, the clays are sandier and contain carbonate concretions and rare well-rolled pebbles. A geologic-geomorphologic cross-section (Fig. 2) shows that the clayey member mapped (*Geological Map* ..., 1963) as the Pliocene unit forms the surface at about 200 m above sea level. This surface is best developed in

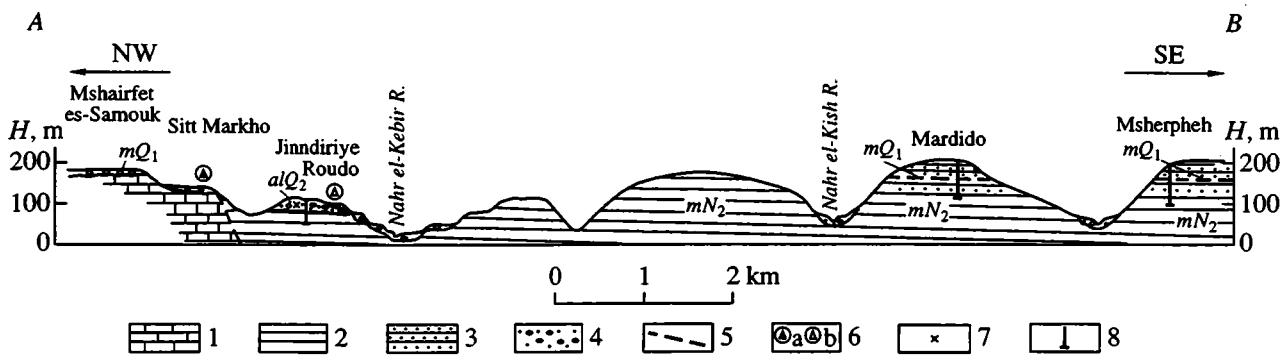

Fig. 2. Geologic-geomorphologic profile along line *A–B*.

1, Paleogene limestones (P); 2, Pliocene marine clays (mN_2); 3, lower Pleistocene marine sandy-clayey deposits (mQ_1); 4, Quaternary marine and alluvial sandstones and pebblestones (*m*, *alQ*); 5, presumed fault; 6, Paleolithic finds: (a) early Acheulean, (b) late Acheulean; 7, finds of larger mammalian bone remains; 8, geologic sequences.

the southernmost part of the Al-Kebir depression. Hypsometrically below this surface, a series of marine terraces and related alluvial terraces is observed along the valley of the Nahr El-Kebir River, draining the depression with the same name.

The highest marine terraces are located at about 170–180 m above sea level. These are the Mshairfet-es-Samouk Formation near Lathakia and the Chaabien Formation on the coast of Lebanon. The top of the terrace near the village of Mshairfet-es-Samouk is composed of pebblestone, 1 to 2 m thick, with well-rolled pebbles. The age of this high terrace is considered to be the early Pleistocene,[1] or pre-Günz (Besancon *et al.*, 1977). The next marine terrace is traced at about 120–130 m above sea level in the area of the mouth of the Nahr El-Kebir River (the Baqsa Formation) and on the Lebanese coast (the Zaqrounien Formation). Geomorphologically, the Baqsa marine terrace corresponds to the alluvial terrace of the Sitt Markho Formation. Alluvial pebblestones of the latter contain artifacts referred to the early Acheulean. Therefore, the Sitt Markho alluvial terrace is attributed to the Günz, whereas the Baqsa marine terrace is referred to the Günz–Mindel (Besancon *et al.*, 1977).

The next stage involving incision of the Nahr El-Kebir River, sedimentation, and formation of the lower coastal line resulted in the appearance of alluvial and marine terraces ranging from 70 to 100 m above sea level. In the Nahr El-Kebir River valley, alluvial terraces are well-developed near the village of Roudo, the alluvial-marine terrace is observed near Khellale, and the marine one is traced near the village of Hennadi (70–80 m above sea level). The first find of bone remains of large mammalia (lower jaw of *Dicerorhinus hemitoechus*, determination by V.Yu. Reshetov) in this region was made in alluvial deposits above Pliocene clays of the Jinndiriye section. These mammalian remains were dated by the Mindel–Riss interval, i.e., the host sediments should be referred to the middle Pleistocene. Artifacts, numerous in the sediments of the 80-m-high terrace near the village of Roudo, where the Jinndiriye terrace is traced, are of the type that is characteristic of the middle–late Acheulean. This also confirms the middle Pleistocene age of this terrace. Geomorphologically, the Roudo and Jinndiriye terraces correspond to the Hennadi marine terrace. The latter is correlative with the Jbailien Formation of the Lebanese coast. In their publications, Sanlaville and Besancon with coauthors refer the Hennadi and Khellale marine terraces to the Mindel–Riss interglacial and the Jinnriye alluvial terrace to the Riss glacial (Besancon *et al.*, 1977; Besancon, 1981; Sanlaville, 1981). This difference in datings of Pleistocene alluvial and marine terraces is not surpising, because in the Pleistocene, according to these authors' concept, marine terraces were formed during interglacial transgressions, while those of alluvial origin are related to glacial regressive phases of the basin development.

Finally, the younger marine terraces, usually considered as Thyrrenian, are located at 20–40 m above sea level. On the Syrian coast, they are developed near the mouth of the Snobar River valley and to the south, near Banyas. Thermoluminescence analysis of marine sands from the 30-m-high marine terrace 3 km south of the village of Snobar, which contains shells of *Acanthocardia tuberculata* (Poirr.), *Cerastoderma* sp., and *Glycymeris* sp. (determinations by A.L. Chepalyga), yielded an age of 101 ± 20 ka.[2] *Strombus bubonis*, a typical Thyrrenian thermophilic form, was found in sediments of the low marine terraces near Banyas. On the Lebanese coast, the terraces of this level are united into the Enfeen Formation, dated (Th/U) 90–100 ka (Sanlaville, 1981). In the Nahr El-Kebir River valley, archeological finds of the middle Paleolithic Age are

[1] The early Pleistocene in the West European stratigraphic scale corresponds to the Eopleistocene in the scale adopted in Russia. In this paper, the West European scale is used.

[2] Age determinations by O.A. Kulikov, Radiochemical Laboratory of Chemistry Faculty, Moscow State University.

associated with the alluvial pebblestones of the 30- to 35-m-high terrace near the village of Esh-Shir, which allows us to refer this terrace to the late Pleistocene.

METHODS OF STUDY AND RESULTS

Various fossil groups, including foraminifers, nannofossils, diatoms, and palynoflora, were studied to obtain data on the biostratigraphy and paleogeography of the late Pliocene–early Pleistocene stage of the Syrian coast. In addition, paleomagnetic measurements of oriented samples were performed, and paleotemperature estimates were made using the Ca/Mg ratio in molluscan shells.

Foraminifers were studied in the Mardido and Al-Quatriya sections. Most samples from the Mardido sequence showed the presence of *Globigerinoides* species, which are indicative of either warm-water or shallow-water conditions. In our case, these indicate shallow-water sedimentation, because benthic forms dominate throughout the entire sequence. In addition, foraminifer assemblages in the upper part of the sequence become impoverished and less abundant. For instance, the lower layers of the section contain 10–12 species, whereas only four to five species occur in its upper part. In the basal layers (0–15 m), the planktonic foraminifer assemblage consists of *Globigerinoides extremus, Gs. Sacculifer, Gs. obliquus, Gs. conglobatus, Gs. amplus, Gs. ruber, Globigerinita glutinata, Globigerina falconensis, G. megastoma, G. tolloides, G. decoraperta, Orbulina universa, Sphaeroidinella dehiscens, Globorotalia humerosa*, and *Globigerinella aequilateralis*. This assemblage correlates the host sediments with the uppermost part of the *Pulleniatina obliquiloculata* Zone and the lowermost part of the *Globorotalia truncatulinoides* Zone of the Atlantic zonal scale (Stainforth *et al.*, 1975). Benthic species *Hyalinea baltica* occurs 13 m above the sequence base.

Similar to the Mardido section, species of *Globigerinoides* genus dominate in the Al-Quatriya sequence. The planktonic foraminifer assemblage correlates the sediments with the interval corresponding to the *Globorotalia tosaensis* Subzone of the *Globorotalia truncatulinoides* Zone. The benthic species of *Hyalinea baltica* is present as well.

Nannofossils were studied in the Jinndiriye, Al-Quatriya, Mardido, and Msherpheh sequences (Fig. 3). Samples were investigated under the light microscope "Jenalumar" with magnification ×1000. Slides were prepared using liquid glass.

A rich and diverse well-preserved nannofossil assemblage was sampled in the Jinndiriye section from the bluish-gray clays (25 m thick) underlying the alluvial sediments. The lower layers of the sequence contain the assemblage of the *Reticulofenestra pseudoumbilica* Zone (NN 15) of Martini's scale. Ten meters above the sequence base, *Discoaster tamalis* appears; it is persistently present from this point upwards. According to the Central Mediterranean scale (Rio *et al.*, 1991), this interval corresponds to the Zanclean–Piacenzian.

Several intervals with nannofossils were established in the Mardido section. Two of them contain a normal marine assemblage that is typical of brackish-water basins. The richest assemblage was recorded at the sequence base (5–10 m). The presence of *Discoaster pentaradiatus* correlates this assemblage with the *Discoaster pentaradiatus* Zone (NN17) of Martini's scale, or with the *Discoaster pentaradiatus* Subzone of the *Discoaster surculus* Zone of the Western Mediterranean stratigraphic scheme (Raffi and Rio, 1979), or with the *Discoaster pentaradiatus* Zone of the chronostratigraphic scale elaborated by Rio *et al.* (1991) for the entire Mediterranean region.

The normal marine nannofossil assemblage in the interval of 65 to 74 m in the Mardido section includes *D. triradiatus* (three-axial variety of *D. brouweri*) and *Gephyrocapsa oceanica* s.l., which appears here for the first time. The first occurrence of the latter is an important stratigraphic marker for defining the Pliocene–Pleistocene boundary. According to Rio, the last occurrence of *D. brouweri* and the first occurrence of *Gephyrocapsa oceanica* s.l. are diachronous events. However, in the Mardido section they are recorded in the same sample, which can be explained either by reworking or by local peculiarities in the nannoplankton habitat. Although the presence of a latent hiatus in the upper part of the sequence cannot be ruled out, this does not depreciate the stratigraphic importance of the facts observed.

Nannofossil distribution in the Msherpheh section is similar to that in the Mardido sequence. As in the latter, the basal layers of the Msherpheh section contain a rich nannofossil assemblage of subtropical affinity. Discoasters are mostly represented by *D. brouweri* and *D. surculus*. Less common are *D. variabilis* and *D.* sp., while *D. pentaradiatus, D. asymmetricus*, and *Ceratolithus rugosus* are absent, implying that the conditions in the Msherpheh locality were more shallow-water than those at the Mardido site. The similar nannofossil composition in both sequences suggests that the lower layers of the Msherpheh section are correlative with the upper part of the *Discoaster surculus* Zone (NN 16) of Martini's scale or with the *Discoaster pentaradiatus* Zone of the Mediterranean scale (Rio *et al.*, 1991). Another stratigraphically important level in the Msherpheh section is in its upper part. As in the Mardido sequence, single specimens of both *D. triradiatus* (*D. brouweri*) and *Gephyrocapsa oceanica* s.l. are recorded in the Msherpheh section, although here they occur at different levels. This fact confirms the suggestion that a latent hiatus is present in the upper part of the former sequence. It should be noted that *Gephyrocapsa oceanica* s.l. from the Msherpheh section is represented by smaller forms than those from the Mardido sequence. The first occurrence of *Gephyrocapsa oceanica* s.l. in the Msherpheh section, recorded 60 m

above its base indicates, that this level is correlative with the Pliocene–Pleistocene boundary.

Gephyrocapsa oceanica s.l. occurs also in the Al-Quatriya section, which refers the host sediments to the Pleistocene or, more exactly, to the lower Pleistocene, because the top layers contain *Helicosphaera sellii*, the index species of the *Helicosphaera sellii* Zone (Gartner, 1977; Rio *et al.*, 1991).

Diatoms were studied in the Mardido section. Only 10 samples of the 27 samples investigated yielded diatom fossils. Their abundance is usually estimated as "single" (1–10 specimens per slide); however, in some samples, they are "common" (500–1000 valves per slide). The total diversity of the diatoms is evident from the 88 species and varieties, including reworked forms. Division Bacillariophyta is represented by two classes: Centrophyceae and Pennatophyceae. The former includes 39 taxa, among which *Thalassiosira* and *Cyclotella* genera are most diverse. The latter consists of 49 species and varieties with *Navicula, Pinnularia*, and *Nitzschia* being more representative. The majority of diatoms in the fossil assemblages are benthic forms (62%), while planktonic species play a subordinate role (38%).

With the respect to salinity, several groups were distinguished. One of the most representative groups consists of marine species. This group makes up 33% of the total assemblage and consists mainly of planktonic neritic species of *Thalassiosira, Porosira, Actinoptychus, Chaetoceros*, and other genera. Marine benthic species are represented by single *Cocconeis costata, Navicula distans*, and *Nitzschia jouseal.* A subordinate share belongs to brackish-marine and brackish-water species (5 and 10%, respectively). They are equally represented by both planktonic and benthic forms. *Cyclotella caspia, Actinocyclus ehrenbergii*, and *Thalassiosira decipiens* are rather common among the latter. A large share among freshwater algae (39.5% of total assemblage) belongs to halophilic species. They make up 14% of the freshwater constituents of the assemblage. The most abundant halophilic species are *Stephanodiscus rotula, S. minutulus, S. hantzschii*, and *Nitzschia frustulum*. Freshwater algae are dominated by indifferent forms that can also tolerate weakly salinated waters (Loseva, 1982).

In general, boreal species prevail (42%) in the diatom assemblage studied, cold-resistant and alpine forms make up 17%, and a relatively large share (41%) belongs to cosmopolitan species. Reworked forms constitute a small share of the assemblage, being represented by Cretaceous and Paleogene species. Based on taxonomic and ecologic analyses of diatom flora from the Mardido section, the following changes in its composition can be noted (from the base upwards).

The assemblage from the lower part of the sequence (16–38 m) is dominated by marine and brackish-marine species. Most abundant among them are planktonic, neritic, and oceanic forms: *Thalassiosira anguste-lineata, T. eccentrica, T. oestrupii*, and *Thalassionema nitzschioides*. Brackish-water forms are recorded; benthic and freshwater species are also common. All this implies the proximity of the shore line, although the assemblage as a whole bears features peculiar to diatom flora of a normal marine basin. Therefore, one can suggest that this assemblage inhabited the shallow-water bay that existed at that time.

Changes in the diatom assemblage occur at the level of 45 m. The diversity of marine species decreases, whereas the share of brackish-water forms increases, although abundances of the latter are low. They include such species as *Actinocyclus ehrenbergii*, typical of littoral areas of sea basins, and *Cyclotella caspia*, common in the Black and Caspian seas. In general, planktonic (both marine and freshwater) species dominate in this assemblage. Species *Stephanodiscus rotula* and *S. minutulus* (almost ubiquists, common in the Black and Mediterranean seas) stand out among the latter group. The presence of freshwater lacustrine species of *Cyclotella* genus together with other planktonic forms as well as the complete absence of benthic forms suggest the existence at that time of a relatively deep brackish-water basin pinched from the open sea.

In the interval of 53–66 m, freshwater, mainly benthic species dominate, indicating the freshening of the paleobasin. Also present are phytons and lacustrine-bog forms characteristic of shallow-water eutrophic basin.

At the level of 68 m, an assemblage similar to that from the interval of 16–38 m occurs. It is dominated by marine neritic and littoral species, among which *Paralia sulcata* is most abundant. Halophilic forms such as *Nitzschia capitellata* and *N. frustulum* prevail among freshwater species. This flora is probably related to a shallow-water marine basin, while the presence of freshwater and benthic species indicates the proximity of the shore line.

The next change in the diatom assemblage composition is recorded in the interval of 87–90 m. The diversity of marine species sharply decreases here, and they are mainly represented by benthic forms. The share of brackish-water benthic species (*Nitzschia sigma* and *N. spectabilis*) increases somewhat. Freshwater benthic species are principal constituents of the assemblage. They are represented by several ecologic groups: bog and lacustrine-bog (*Eunotia lunaris, Pinnularia subcapitata*, and *P. viridis* var. *leptogongyla*), soil (*Navicula mutica* var. *cohnii, Pinnularia borealis* var. *brevicostata*, and *Hantzschia amphioxys*), and fluvial (*Navicula cryptocephala* and *N. minima*). This composition suggests that the assemblage habitat was a shallow-water basin, strongly freshened but still connected with the sea. This could be an estuary, locally swamped and periodically drained.

The composition and morphologic features of the diatom flora from the lower part of the considered section suggests its Pliocene age. This is indicated by the

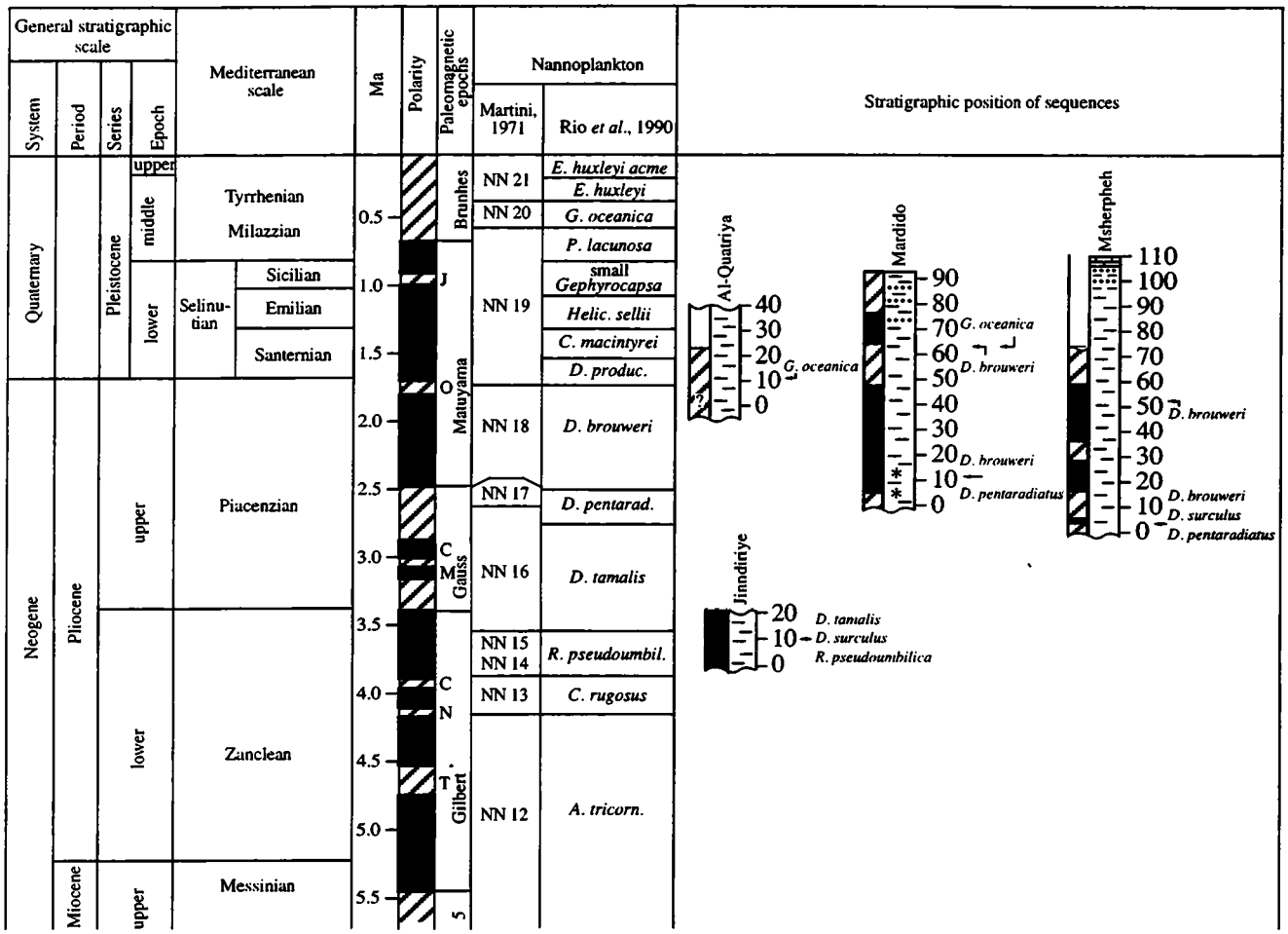

Fig. 3. Stratigraphic correlation of Pliocene–Quaternary marine deposits in western Syria and their paleogeographic characteristics.

presence of a large group of extinct species, which make up 16% of total assemblage. These are mainly species that only existed in the Neogene, such as *Thalassiosira manifesta, Cyclotella omarensis, C. perforata*, and *Rhaphoneis* cf. *maeotica*.

Palynoflora was studied in the Jinndiriye, Al-Quatriya, and Mardido sequences. The Pliocene part of the first section underlying the alluvial sediments contains palynological spectra dominated by pollen of the arboreal-shrub group (up to 80%). Most abundant are pollen of *Pinus* sg. *Haploxylon* (up to 90%), *Pinus* sg. *Diploxylon* (up to 6%), *Picea* (up to 12%), *Quercus* (4–8%), and *Oleaceae* (2–6%). Pollen of *Abies, Cedrus, Betula, Ostrya, Alnus, Fraxinus, Corylus, Juglans, Salix*, and *Palma* are present as single grains. Herbaceous pollen are mainly represented by Chenopodiaceae (10–20%), Compositae (4–15%), and Gramineae (8–20%). Although the share of spores in the spectra is insignificant, these constituents are diverse and represented by Polypodiaceae, *Cystopteris, Anemia, Licopodium, Selaginella, Botrichium*, and *Osmunda*. This composition of the palynospectra allows us to suggest a wide distribution of coniferous (*Picea* and *Pinus*) forests in the mountain areas. The arboreal vegetation also included *Abies, Cedrus*, and *Podocarpus*. Piedmonts were covered by mixed coniferous and broad-leaved forests, while plains were probably occupied by maquis with an admixture of steppe vegetation. The climate at that time was much more humid and somewhat cooler than the modern one.

The composition of the palynological spectra in the Mardido section shows variations in ratio of arboreal and herbaceous pollen against a background of predominating arboreal forms (more than 50%).

Eight palynocomplexes united into three palynoflora are established in the interval from 7 to 86 m. Samples from the interval of 7–30 m (palynoflora I) are characterized by a prevalence of arboreal pollen of *Cedrus, Picea, Pinus*, and *Abies*. Taxodiaceae, *Tsuga, Podocarpus, Betula, Fagus*, and *Eucommia* are also present. The herbaceous group is dominated by Gramineae, Compositae, and Rhodontheae. These palynocomplexes reflect the development of mixed

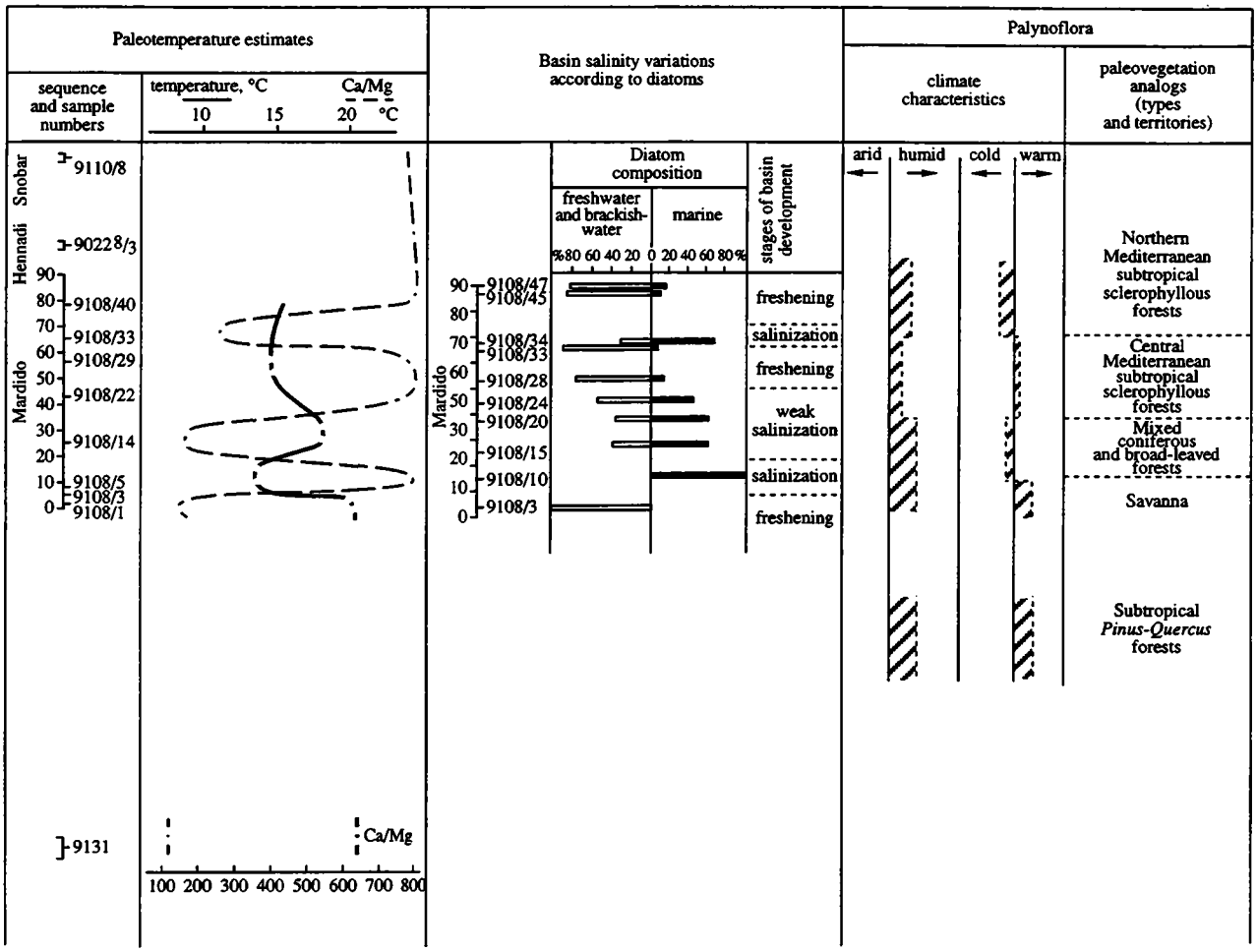

Fig. 3. (Contd.)

coniferous and broad-leaved forests in the piedmont and mountainous ares. Plains were covered by evergreen sclerophyllous forests with steppe patches.

The interval from 30 to 60 m (playnoflora II) shows frequent variations in pollen composition. Pines and herbs dominate. Arboreal pollen of *Abies, Picea, Cedrus, Liquidambar, Juglans,* Oleaceae, and *Quercus* are present in equally insignificant quantities. *Cistus* and *Ceratonia* occur as single grains. Herbs are represented by pollen of Chenopodiaceae, *Ephedra*, Cruciferae, Leguminesae, and Umbeliferae. The mountains were covered by typical Mediterranean evergreen sclerophyllous forests, and plains were occupied by maquis vegetation.

Palynoflora III, characteristic of the interval from 60 to 86 m, shows an increased content of arboreal pollen: *Pinus* sect. *Strobus, Cedrus, Picea*, and *Abies*. Pines dominate; *Tilia* and *Magnolia* occur as single grains. Lowlands and plains were covered by sclerophyllous forests and shrubs, whereas highlands were occupied by *Pinus* and *Cedrus* with an admixture of spruce and fir. Mountainous forests of that time were composed of deciduous *Tilia, Alder, Betula*, and *Fagus*.

In the Al-Quatriya section, palynospectra are dominated by arboreal pollen (up to 40–50%), among which *Pinus* (38%), Oleaceae (up to 28%), *Quercus* (27%), and *Ulmus* (9%) are most common. Pollen of *Abies, Cedrus, Platicarya, Alnus, Tilia, Ostrya*, and *Liquidambar* also occur. Herbaceous pollen are represented by Taraxacaceae, Chenopodiaceae, Asteraceae, Cruciferae, Gramineae, and Ranunculaceae. Single pollen grains of *Anemia, Ophyoglossum, Lycopodium*, Polypodiaceae, and *Pteris* were also recorded in this section (Simakova, 1993). Palynologic data from the Al-Quatriya sequence allows us to suggest that mixed coniferous and broad-leaved forests grew in the mountains, maquis vegetation covered piedmonts and areas along water streams, and steppe vegetation occupied plains.

Paleomagnetic measurements were performed in the Jinndiriye, Al-Quatriya, Mardido, and Msherpheh sections. Approximately 120 oriented samples were collected from deposits with a total thickness of 250 m. Sampling intervals were approximately every 2 m. Two

cubes with an edge length of 2 cm were cut from each sample. In the laboratory, they were subjected to step heating up to 400°C. Heating to higher temperatures was unreasonable, because in the 350–400°C range, strong additional magnetization had already appeared. Component analysis revealed both recent viscous magnetization and an ancient one. The latter became apparent in the temperature range from 200 to 330°C. Measurements were performed using a rock-generator JR-4. Values of natural remanent magnetization in the Pliocene samples varied from 1 to 10 nT during the first measurements. After demagnetization, I n decreased and varied from 0.1 nT to 1.0 nT in the temperature range from 200 to 330°C. Because of the strong additional magnetization that resulted from laboratory sample treatment, the results of demagnetization cannot be considered as completely reliable; however, the sign of magnetization was unambigously determined in most cases. Paleomagnetic studies revealed reversed polarity in the lower (Pliocene) part of the Jinndiriye section. The Mardido and Msherpheh sequences overlap each other to some extent; however, the former is more complete in its upper part, while the latter is more complete in its lower part. The basal layers of both sections are characterized by normal polarity. Upwards, a wide zone of reversed polarity is observed with several normally magnetized layers. The Al-Quatriya sequence is characterized by normal polarity, although no reliable results were obtained in its upper part.

In general, paleomagnetic measurement results combined with the available geologic and paleontologic data can be interpreted in the following way. The Jinndiriye section corresponds, evidently, to the upper part of the Gilbert Chron. The normally magnetized lower part of the Mardido and Msherpheh sections correlates with the Gauss Chron, and the thick member of reversely magnetized rocks composing their upper part undoubtedly corresponds to the Matuyama Chron. Normally magnetized layers in the upper part of these sequences (inside the wide zone of reversed polarity) and in the lower part of the Al-Quatriya section can probably be correlated with the Olduvai Event. Thus, the sequences studied correspond to the time interval of at least 3.8 to 1.6–1.0 Ma on the paleomagnetic scale (Fig. 3).

Paleotemperature estimates were made using the magnesium method applied to marine bivalve shells collected in a layer-by-layer manner from the Mardido section and from separate intervals in the sections of the Hennadi and Snobar marine terraces (Fig. 3). In addition, large *Ostrea* shells from the Pliocene basal layers that outcrop 15 km to the north of Jisr Ash-Shugur were analyzed. Paleotemperature estimates are based on the Ca/Mg ratio in shells of marine invertebrates, depending on the average temperatures of the environments. It is well-known fact now that, despite the dependence of the Ca/Mg ratio on the temperature regime, it is also predetermined by taxonomy. Organisms of different taxonomic groups evolving under the same temperature environments accumulate different quantities of magnium and strontium. Therefore, paleotemperature scales for separate marine invertebrates were elaborated (Slavin and Yasamanov, 1982).

In this paper, the magnesium method is used to analyze bivalve mollusks that inhabite marine basins with salinity that is close to normal and that have shells of an essentially calcite composition. The Ca/Mg ratio was interpreted using the paleotemperature scale calculated by this ratio in recent *Ostrea, Chlamys, Panopea, Tapes*, and *Pecten* (Slavin and Yasamanov, 1982; Yasamanov, 1989).

The Ca and Mg contents and their ratio were measured in 12 samples of bivalve mollusk shells (*Pecten, Cardium,* and *Ostrea*). Based on the Ca/Mg ratio in large *Ostrea* shells from the basal layers of the lower Pliocene, the annual average temperature or average temperature of habitat environments of these organisms was estimated to have been about 20°C. Similar temperature values (20–22°C) were characteristic of the middle late Pliocene. This agrees with the palynological data suggesting that the territory under consideration was covered by a savanna at that time. Then, at 2.5–2.4 Ma, the temperature, according to paleotemperature estimates, fell to 13–15°C. This was a time of development of mixed coniferous and broad-leaved forests. At the end of the Pliocene, this cool period was followed by a warming, when maximum temperature values reached 18°C. As a result, mixed coniferous and broad-leaved forests transformed into forests of the Central Mediterranean type (Fig. 3). In the early Pleistocene, temperature fell again to 15°C.

It should be noted that not all of the measurements performed gave unambiguous results. In some cases, the considered variations in shell composition can be attributed to the influence of salinity changes. This suggestion seems to be valid because the studied bivalve mollusks inhabited shallow-water bays that periodically became isolated and alternately experienced salination and freshening. Humidity could also have contributed significantly to these changes.

STRATIGRAPHY

The results of biostratigraphic and paleomagnetic studies show that the considered sequences encompass different intervals of the Pliocene and lower Pleistocene. The data obtained allow us to establish the stratigraphic succession and define the position of stratigraphic boundaries. The oldest Pliocene sediments are recovered in the Jindiriye section (below alluvial pebblestones). The nannofossil assemblage of the NN 15 and the NN 16 zones of Martini's scale recorded in these layers as well as paleomagnetic data indicating their reversed polarity correlate the respec-

tive interval with the Gilbert Chron. In addition, taking into account the position of the relatively narrow NN 15 and NN 16 zones, there is reason to believe that this sequence corresponds to the uppermost lower Pliocene.

Judging from paleomagnetic data and the results of nannofossil studies, the Mardido and Msherpheh sections encompass practically the same age interval, i.e., the second half of the late Pliocene to the early Pleistocene, or approximately from 2.8 Ma to, probably, 1.0 Ma. A diverse nannofossil assemblage with numerous *D. brouweri, D. pentaradiatus, C. macintyrei, P. lacunosa*, and other forms correlates the normally magnetized interval in the upper part of these sequences with the upper part of the Gauss Chron. Upward in the section, the composition of nannofossils becomes impoverished. *D. brouweri* occurs from the sequence base to the level of 65 m, where the first occurrence of *Gephyrocapsa oceanica* s.l. is also recorded. According to paleomagnetic measurements, this level corresponds to the upper part of the short normally magnetized interval, which is identified with the Olduvai Event. In the Msherpheh sequence, the first occurrence of *Gephyrocapsa oceanica* s.l. is recorded at approximately the same level (60 m above the base) as that in the Mardido section. The normally magnetized interval occurs in the upper part of the sandy-clayey member of the Mardido section, which can probably be identified with the Jaramillo Event. Nannofossil and paleomagnetic data make it possible to define the position of the Pliocene–Pleistocene boundary in the Mardido and Msherpheh marine sequences. This boundary, associated with the top of the Olduvai Event, occurs 30–40 m below the top of the sandy-clayey member outcropping in these sequences (Figs. 2 and 3). Lithologically, the upper 30- to 40-m-thick member of marine sediments is enriched in sandy material. As for the stratigraphic correlation of the uppermost part of the sequence above the proposed Pliocene–Pleistocene boundary, the poor paleontologic characteristic of this interval should be taken into consideration. It seems reasonable to correlate the post-Pliocene part in the Mardido and Msherpheh sections with the Selinuntian, which corresponds to the interval from 1.65 to 0.8 Ma in the stratigraphic scale (Rio *et al.*, 1991). Marine deposits overlaying the Pliocene rocks form a terrace at 200 m above sea level. This terrace is widely developed in the southwestern part of the Al-Kebir depression and can be correlated with the Calabrian one. In this context, the Calabrian is considered solely in a geomorphological context, because in the stratotype section, (Calabria, Italy) the Calabrian corresponds not to the entire lower Pleistocene, but to its smaller upper part (Rio *et al.*, 1991). In the coastal area near Lathakia, all young marine and alluvial terraces with an age younger than 0.9–1.0 Ma are inset in the Calabrian terrace.

PALEOGEOGRAPHY

Geomorphologically, the Al-Kebir depression with the Pliocene–Quaternary sediment infilling belongs to the Eastern Levantine basin that is more than 300 km long and 30–40 wide, stretching along the Eastern Mediterranean coast. The average thickness of Pliocene–Pleistocene sediments in the submarine part of the basin is up to 1–1.5 km (Chumakov and Yutsis, 1981). Available data on the marine fauna and nannoflora development show that the maximum of the Pliocene transgression in the Eastern Mediterranean occurred in the early Pliocene and at the beginning of the late Pliocene. The terminal Pliocene was marked by a gradual regression. For instance, the Jinndiriye section corresponding to the Zanclean–Piacencian boundary interval is characterized by an abundant and diverse nannofossil assemblage, whereas later, beginning approximately at 2.5 Ma, the latter becomes impoverished. Similar distribution patterns are characteristic of foraminifer and diatom assemblages. The marine basin on the Syrian coast was characterized in the terminal Pliocene by shallow depths. This is evidenced by a foraminifer assemblage in which benthic forms dominate. Diatoms also show that marine environments in the terminal Pliocene were gradually replaced by those of a marine bay or estuary (Fig. 3).

The transition between the Pliocene and the lower Pleistocene in the studied sequences is relatively gradual, although the higher content of sandy material in the lower Pleistocene sediments distinguishes them from the more clayey upper Pliocene deposits. The early Pleistocene was marked by more shallow-water depositional environments with a significant input of terrigenous material and the swamping of some coastal areas associated with the appearance of obvious subaerial settings. The latter is evidenced by soil reworking in certain intervals within the upper part of marine members, as is recognized, for instance, in the Al-Quatriya and Mardido sequences.

Changes in paleofloral assemblages reconstructed by palynological data revealed the general trend and main stages of climatic fluctuations in the Pliocene through the early Pleistocene.

At the end of the early Pliocene, when coniferous vegetation had a maximum (for the studied time interval) distribution, the climate was humid and relatively cold. At approximately 3.4 Ma, warming commenced against a background of high humidity, which is evidenced by a decreased share of coniferous in the arboreal assemblage and an inreased role of thermophilic *Quercus, Carpinus*, and Oleaceae.

Cooling at 2.4 Ma, against a background of a relatively humid climate, resulted in the wide distribution of mixed coniferous and broad-leaved forests with *Tsuga, Fagus*, and *Liquidambar*. This cooling was recorded on the territory of Israel (Horowitz, 1990) as well as in the northern Mediterranean, where it is cor-

relared with a Pretiglian cooling associated with climate aridization (Suc and Zagwijn, 1983; Suc, 1984).

Beginning in the terminal Pliocene, polycyclic climatic variations (arid–humid) become well-pronounced. Periods of aridization were marked by a sharp decrease in the abundance of mixed coniferous and broad-leaved forests, whereas maquis and steppe vegetation became widespread. Evergreen sclerophyllous forests of the Mediterranean type were the principal type of vegetation in these time intervals.

At 1.8 Ma, cooling with insignificant climate humidization occurred in the considered region, which was then followed by a period of warming recorded in the northern Mediterranean as well (Gillet *et al.*, 1978).

In general, the late Pliocene–early Pleistocene was characterized by polycyclic climate variations that occurred against a background of directed aridization and relative cooling.

Decreases in the paleobasin water temperature at 2.4 Ma, in the terminal Pliocene, and at the beginning of the Pleistocene probably reflect the reorganization of the thermal regime in the Eastern Miditerranean in response to general cooling on the continent.

CONCLUSION

(1) The stratigraphic subdivision of the upper Pliocene deposits in western Syria was specified and the position of the Pliocene–Quaternary boundary was outlined. In the studied sequences, this boundary is well-defined by the last occurrence of *D. brouweri* and the first occurrence of *G. oceanica* at the level of the normal polarity event correlated with the paleomagnetic Olduvai Event.

(2) The terminal phase of transgression on the western Syrian coast can be dated approximately at 1.0 Ma. This event corresponds to the Sicilian–Calabrian. The maximum of the Pliocene transgression occurred in the early Pliocene and at the beginning of the late Pliocene. In the early Pleistocene, a shallow-water marine basin existed in western Syria.

(3) Considering new data on the stratigraphy of upper Pliocene–Pleistocene deposits as well as paleontologic and archeologic remains from alluvial terraces in the low reaches of the Nahr El-Kebir River valley, one can hypothesize that the age of the terraces at a height interval of 100–200 m above sea level ranges from 0.6 to 1.0 Ma.

(4) The late Pliocene–early Pleistocene was a time of polycyclic climate oscillations that occurred against a background of progressive aridization and cooling. According to palynologic data and paleotemperature estimates on the base of Ca/Mg ratios in marine mollusk shells, the cooling event at 2.4 Ma corresponds to the Pretiglian cooling in Western Europe.

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Stratigraphy and Volcanism in the Early Proterozoic Pechenga Structure (Kola Peninsula)

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Abstract—The Pechenga structure comprising sequences, which serve as the reference for the northeastern Baltic Shield, includes the northern and southern structural zones. Both zones are composed of sedimentary-volcanogenic rocks of the Karelian Complex (the Lower Karelian Eratem) metamorphozed under conditions of prehnite to pumpellyite facies. Based on published and new original data, the following formations are proposed to be defined in the northern zone (from the base upward): the Televi, Majärvi, Kuvernerinjoki, Pirttijärvi, Orshoaijvi, Luchlompolo, Zapolyarnyi, Zhdanov, Lammas, Matert, and Suppvaara formations correlative to the Sariolian, Jatulian, and Ludicovian (Ludian) of Karelia. In the southern zone, the Karelian complex is divided into the Kallojaur, Bragino, and Mennel' formations and Kaplya and Kasesjoki rock associations, which are referred to the Kalevian. The paleovolcanologic, formational, and geochemical analyses enabled the recognition of the basaltic andesite, trachybasalt, tholeiite, ferropicrite, picrite-basalt-andesite, and andesite-dacite-rhyolite volcanogenic formations, all formed during six megacycles of magmatism in the interval of 2350–1850 Ma. Metamorphism culminated at 1750–1700 Ma. The structural patterns of the sequences and geochemical and isotopic data agree with the model of intracontinental riftogenesis in the Pechenga-Varzuga belt, which experienced a complete cycle of development during the Early Proterozoic.

Key words: *Baltic Shield, Pechenga, Lower Proterozoic, stratigraphy, volcanism, Karelides, geochemistry, geochronology*

INTRODUCTION

The Pechenga structure in the northeastern Baltic (Fennoscandia) Shield attracted the attention of geologists for a long time because of the ore deposits (Cu–Ni-sulfide) discovered there. Moreover, this area provided key data for understanding the Early Proterozoic history of the continental crust formation in the Kola province (Zagorodnyi and Radchenko, 1988). In 1987–1994, complex geological–geophysical investigations of this structure were carried out in the scope of the state program *Geodynamic Project-50* and the Russian–Norwegian *Northern Project*. These works resulted in: (1) revision and clarification of Karelian complex stratigraphy; (2) compilation of the new geological map of the structure with a 1 : 50 000 scale; (3) subdivision of volcanogenic deposits; (4) more accurate determination of the geological relations between rocks of variable composition, including those between the volcanogenic and spatially associated intrusive rocks; (5) discrimination of volcanogenic associations of different geneses; and (6) geochronological dating of magmatic rocks using the Sm–Nd, Rb–Sr, Re–Os, Pb–Pb, and U–Pb methods. To understand the rock genesis and trace their evolution throughout the sequence, we studied them using the atomic absorption, x-ray fluorescence, and neutron activation methods.

GENERAL GEOLOGY

The Pechenga structure (Fig. 1) represents a part of the Early Proterozoic Pechenga-Varzuga belt, which stretches with local gaps along a distance of 600 km from the settlement of Polmak via the Pasvik River, northern Norway, to the Pechenga River and Keulik tundra, and further to the southeast of the Lake Imandra via the basins of the Varzuga and Strel'na rivers to the Sosnovka River in Russia (the White Sea neck) (Zagorodnyi and Radchenko, 1988; Smol'kin, 1993). The area is more than 2000 km^2. It is traditionally subdivided into two, northern and southern structural zones different in rock occurrence (flat monoclinal and steep isoclinal, respectively), grade of metamorphism (mostly greenschist and amphibolite facies, respectively), rock associations, and basement composition. The structural and stratigraphic relations of these two zones were under intense discussion (Zagorodnyi *et al.*, 1964; Zagorodnyi and Radchenko, 1988; Gilyarova, 1976; Polyak, 1968; Predovskii *et al.*, 1974, 1987; Kozlovskii *et al.*, 1984). The works performed in accordance with the aforementioned project resulted in the elaboration of a new stratigraphic scheme, which takes into account both published data, and original materials (table).

According to this scheme, the northern zone is composed of the Karelian Complex, 2.4 to 1.9 Ga in age, whereas the rocks of the same complex composing the

Scheme of stratigraphic subdivisions in the northern and southern zones of the Pechenga structure

Eon	Complex	Superhorizon	Lithostratigraphic unit	Volcanic, sedimentary formation; intrusive complex
			Southern Pechenga zone	
Lower Proterozoic	Karelian	Kalevian	*Porojärvi Group*[6]	
			Kasesjoki association[2]	Conglomerate-shaly, molassoid fm.
				Dacite-rhyolite subvolcanic complex
			Kaplya association[6]	Andesite-dacite-rhyolite fm
			Ansemjoki Group[6]	
			Mennel' association[6]	Picrite-basalt-andesite fm
			Bragino Formation[6]	
			Kallojaur Formation[5]	Tuff-shaly fm.
			Nortern Pechenga zone	
		Ludian	*Pilgujärvi Group*[1]	
				Gabbro-wherlite complex
			Suppvaara Formation[6]	
			Matert Formation[3]	(Ferro)picrite-basalt fm.
			Lammas Formation[3]	
			(Upper productive)	Terrigenous (shaly),
			Zhdanov Formation[3]	
			(Lower productive)	Flyschoid fm.
			Kolosjoki Group[1]	
		Upper Jatulian	Zapolyarnyi Formation[4]	Tholeitic basalt fm.
			Luchlompolo Formation[3]	Terrigenous-tuff-carbonate fm.
			Kuetsjärvi Group[1]	
		Lower Jatulian	Orshoajvi Formation[6]	Trachybasalt fm.
			Pirttijärvi Formation[4]	
			Kuvernerinjoki Formation[3]	Quartzite-carbonate fm.
			Akhmalakhti Group[1]	
		Sariolian	Majärvi Formation[4]	Andesite-basalt fm.
			Televi Formation[3]	Basal, conglomerate fm.
			Keulik-Kenirim (eastern Pechenga area);	
		Sumian	Keulik Sequence[2]	Basalt fm.
				Peridotite-pyroxenite-gabbronorite complex
2500 ± 50				
Upper Archean		Loppian		

Note: The names and volumes of groups and formations are given after: 1, Zagorodnyi *et al.*, 1964; 2, Polyak, 1968; 3, Kolesnikov *et al.*, 1972 (unpublished data) and Chalykh *et al.*, 1973 (unpublished data); 4, V.S. Lanev *et al.*, in Kozlovskii *et al.*, 1984; 5, Golubev *et al.*, 1984; 6, original works.

southern zone are dated as 1.85 Ga old. In the Precambrian stratigraphic scheme of Russia, which was adopted in Ufa (Semikhatov *et al.*, 1991), these deposits belong to the Lower Proterozoic (Karelian) and should be referred to the following units distinguished in Karelides of the Baltic Shield: the Sariolian, Jatulian, and Ludicovian in the northern zone, and the Kalevian in the southern zone. The upper age limit of the succession is defined by synchronous events (1.75–1.70 Ga) of Svecofennian orogeny and metamorphism. In accordance with the requirements of the modern stratigraphic code (*Stratigraficheskii kodeks*, 1992), the Karelian Complex of the Pechenga structure was subdivided into groups, subgroups, and formations. The principal taxonomic unit is represented by the formation which can be shown in maps as a separate geological body with definite stratigraphic contacts. When such boundaries are unclear, or of the tectonic type, the unit is ranked as a rock association. Formations and subgroups are united into the groups when they are units of a single transgressive-regressive cycle (for sedimentary deposits), or a single petrochemical series (for

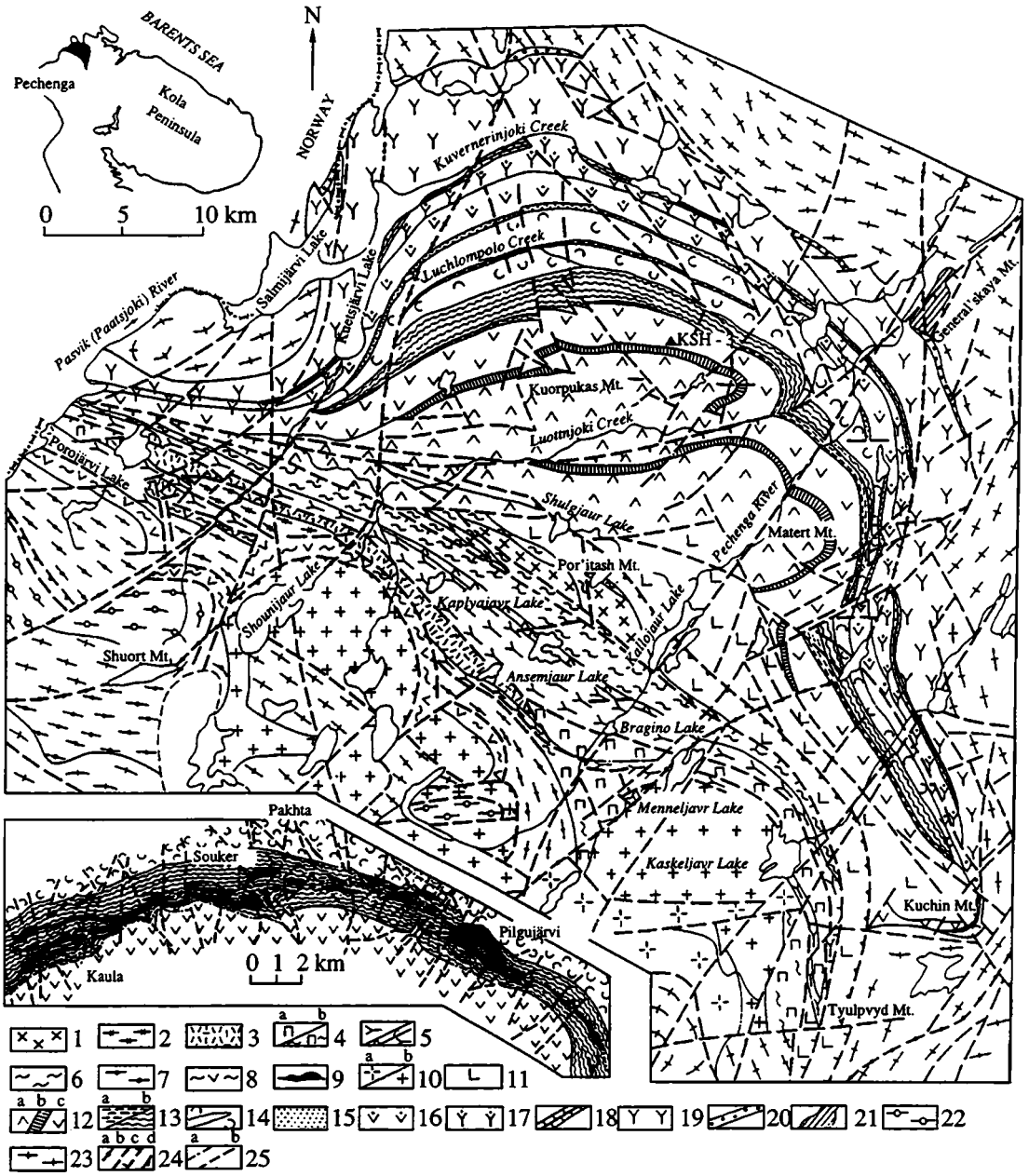

Fig. 1. Geological sketch map of the Pechenga structure compiled on the basis of map at the scale 1 : 50000. Inset shows position of the Pechenga ore field.

1, dacites and rhyolites (I and II phases, respectively), the Por'itash subvolcanic complex; 2, tuff conglomerates, gravelstones, and sandstones, the Kassesjoki rock association; 3, andesites, dacites, rhyolites, and their tuffs, the Kaplya rock association; 4, picrites, basalts and their tuffs, the Mennel' association; 5, basalts, andesite-basalts, andesites, tuffs, and sandstones, the Bragino Formation; 6, sandstones, siltstones, tuffs, and siliciliths, the Kallojaur Formation; 7, biotite, bimicaceous, and garnet-micaceous gneisses, the Tal'ino rock sequence; 8, undivided schistose amphibolites; 9, gabbro, clinopyroxenites, wherlites, and serpentinites, the Pechenga intrusive complex; 10, diorites, granodiorites (a) granites, and quartzose granites (b) of dome structures (Shuoni, Kaskel', and others); 11, basalts, the Suppvaara Formation; 12, tholeiitic basalts, ferropicrites, and their tuffs, siliceous shales of the upper (a), middle (b), and lower (c) members of the Matert Formation; 13, picrite and basalt tuffs of the Lammas Formation (a), sandstones, siltstones, sulfide-carbonaceous shales, and phosphorites of the Zhdanov Formation (b); 14, tholeiitic basalts and shales, the Zapolyarnyi Formation; 15, sandstones, gravelstones, and dolomites, the Luchlompolo Formation; 16, basalts and trachybasalts, Orshoaivi Formation; 17, trachybasalts, basaltic trachyandesite, trachyandesite-dacites, and trachytes, the Pirttijärvi Formation; 18, quartzites and dolomites, the Kuvernerinjoki Formation; 19, basaltic andesites and dacites, the Majärvi Formation; 20, basal conglomerates, gravelstones, and sandstones, the Televi Formation; 21, gabbronorites, bronzite, and anorthosites of the General'skaya Mt. massif; 22, 23, the Upper Archean complex: 22, high-Al gneisses, 23, gneisses, amphibolites, and migmatites; 24, tectonic dislocations: long-active, (a) synsedimentary; (b) steep faults, (c) upthrusts and shifts, (d) and overthrusts; 25, (a) upthrusts, (b) reversed and strik-slip faults in the Pechenga ore field.

volcanics). In the northern zone, the former Akhmalakhti, Kuetsjärvi, Kolosjoki, and Pilgujärvi formations (Zagorodnyi *et al.*, 1964) were ranked as groups with the respective names. In the proposed scheme, each of them incorporates from two to four sedimentary and volcanogenic formations (table). Five sedimentary and sedimentary-volcanogenic formations and rock associations of the southern zone compose the Ansemjoki and Porojärvi groups. In most cases, we retained the former names of the formations taking into consideration their validity. Sometimes, mainly in the southern zone, the cardinally new data on sequence structures and rock relationships served a basis for introducing new stratigraphic units.

STRATIGRAPHY AND VOLCANISM OF THE KARELIAN COMPLEX IN THE NORTHERN ZONE

The Archean basement complex. The Kola superdeep hole (KSH-3) penetrated the contact between the rocks of the Karelian Complex and Archean basement at the depth 6842 m; the contact shows signs of angular unconformity (Kozlovskii *et al.*, 1984). The heterogeneous basement consists of various gneisses, amphibolites, jaspillites, and migmatite granites. These rocks were traditionally referred to the Kola Group of the Upper Archean. In V.R. Vetrin's opinion (cited from Mitrofanov *et al.*, 1991), the basement is mainly composed of gray gneisses which are exposed in northern Norway. In general, the basement rocks form isoclinal folds, and mostly have a northeastern strike. Judging from zircon datings, the oldest rocks are represented by tonalite gneisses penetrated by KSH-3, which crop out in the northern framing of the structure and in northern Norway (Bibikova *et al.*, 1993a; Levchenkov *et al.*, 1993). Their age range is 2930–2903 Ma. Granodiorites from the Holmvatn area (northern Norway) and Tulom massiff (south of Murmansk) that intrude the gneisses of the Kola Group show the U–Pb age of 2762–2729 Ma, whereas plagiogranites in KSH-3 and the Zapolayarnyi area (the Pit'evoe Lake) are 2752 ± 13 Ma and 2753 ± 22 Ma old, respectively (Mitrofanov *et al.*, 1991; Bibikova *et al.*, 1993b; Levchenkov *et al.*, 1993). The rocks of the Archean basement and northern framing of the structure repeatedly experienced metamorphism and granitization 2.5, 2.45–2.30, 2.25–2.20, 2.10–1.90, and 1.82–1.78 Ga ago that was related to the multiple activation of tectonic movements and basic magmatism during the Proterozoic (Duk *et al.*, 1989; Mitrofanov *et al.*, 1991; Bibikova *et al.*, 1993b).

The Karelian Complex

The Sumian. In the Pechenga area proper, no Sumian rocks have been distinguished thus far. Southward, in the distribution area of Archean gneisses, the amphibolite rock association is exposed in elevations of the Keulik, Kenirim, and Podvyd tundras. Amphibolites were intruded by serpentinites and influenced by the Litsa-Araguba granitoids. The sequences are dominated by metamorphozed basaltic lavas with subordinate tuffs and tuffites. Previously, it was established that these rocks are close in sequence structure and geochemical features to amphibolites of the Purnach and Kukshino formations (Smol'kin, 1992) composing the hypostratotype of the Sumi Superhorizon in the Kola province (*Problemy* ..., 1976). This suggests that stratified Sumian rocks are present in the Pechenga area.

In the northern framing of the Pechenga structure, there are many layered intrusions composed of lherzolite-bronzite-gabbronorite-gabbro-diorite associations (the General'skaya Mt., Karicjavr, Saken, and other localities), and coeval gabbronorite dykes. The Sm–Nd age of intrusive rocks from the General'skaya Mt. locality is 2453 ± 42 Ma (Balashov *et al.*, 1993a), whereas that determined by the U–Pb method is 2505 ± 2 Ma (Bayanova *et al.*, 1995). Metamorphism of the layered intrusive rocks was accompanied by granite injections. The Pb–Pb age of one of the granite veins in the Karicjavr locality is 1939 ± 60 Ma. In the area of the General'skaya Mt. southern piedmont, the borehole penetrated eluvial gabbronorite breccia, which is overlain with angular unconformity by basal terrigenous rocks of the Pechenga sequence.

The Sariolian. In the northern Pechenga zone, the Sariolian deposits include the basal horizon of polymictic conglomerates, arkoses of the Televi Formation, and volcanogenic rocks of the Majärvi Formation. The total thickness of the Sariolian sequence varies from 1600 m in the central part of the zone, to 100–200 m on its flanks. These two formations are united into the Akhmalakhti Group, which is bordered at the top by the regional weathering surface. The most complete sections of the Televi Formation is exposed in the northeastern extremity of the Salmijärvi Lake (the Kumpula Peninsula, Televi), and along the Malaya Pechenga eastern riverbank (south of the Luostari railway station) (Zagorodnyi *et al.*, 1964; Predovskii *et al.*, 1974). Formerly, before the construction of the Borisoglebsk hydroelectric station, the 180 m-thick succession of the formation could be observed in the Kumpula Peninsula area. Three major rhythms of approximately equal thickness were recognized in this succession. At the base of every rhythm, coarse cross-bedded units of conglomerates and gravelstones occur. The clast composition is dominated by plagiogranites, granodiorites, and granite-gneisses cemented by gravely material. The middle and upper parts of the rhythms are composed of gravely and coarse-grained psammites. South of the Luostari railway station, the thickness of the formation increases to 250 m. It gradually dips southeastward under volcanics, and rests with angular unconformity upon the Archean gneisses dissected by numerous plagiogranite and, less commonly, plagioclase-microcline granite veins. The basal layers of the sequence contain pebble-gravely conglomerates

cemented by dark gray sandstone, whereas its main part represents the rhythmical alternation of conglomerates, cross-bedded gravelstones, and psammites with an admixture (in the upper part) of tuffaceous material. Judging from the orientation of the cross-bedded series and clast composition, the provenance was located to the north, in an area of physical weathering of the granite-gneiss basement. V.Z. Negrutsa (cited from Smol'kin *et al.*, 1995) believes that the clastic material was transported by seasonal flows, which discharged in both the coastal zone with active hydrodynamics, and in the more distant areas of the basin. The formation of terrigenous deposits proceeded under transgressive conditions, as is evident from the facies distribution.

New data on the Televi Formation structure were obtained in the adjacent territory of northern Norway. In the Neverskrug area, the basal horizon of the formation rests upon the weathering surface of Archean plagioclase-microcline granites. The outcrops exhibit a gradual transition from massive to fractured granites, then to coarse cemented gruss and gruss with poorly rounded pebbles and, finally, to coarse cross-bedded conglomerates and gravelstones with clear rhythms. The thicknesses of the rhythms vary from 100 to 250 m; the total thickness of the formation is sometimes up to 1.5 km. The lower part of the rhythms is generally dominated by coarse, poorly sorted boulder-pebble material, while gravely material with clear bedding prevails in their middle and upper parts. Boulders and pebbles are represented by plagioclase-microcline granite, granodiorite, amphibolite, bimicaceous and garnet-biotite gneiss, and jaspillite. The latter is similar to that from the nearby Sudvaranger jaspillite deposit. Based on the succession structure, these deposits should be referred to the channel and deltaic facies. The material was transported in a north–south direction. Eastward, near the Russian–Norwegian boundary, the basal sedimentary horizon sharply wedges out, and volcanogenic rocks directly overlie the Archaean granite-gneiss basement. The lowest lava flow is composed of basaltic andesite with porphyritic crystal clusters.

The basal horizon of the overlying Majärvi Formation comprises the tuff layer and thin flows of low-titanium and picritic basalts, whereas the main part of the succession is composed of basaltic andesite and, less commonly, andesite and andesite-dacite flows 0.5 to 15–25 m thick. Judging from the successions which are penetrated by the IX (Fig. 2) and KSH-3 boreholes (Kozlovskii *et al.*, 1984) or exposed in outcrops, normal to picritic basalts are mainly characteristic of the lower and, less commonly, middle parts of the unit, whereas acid volcanics are mostly confined to their upper part. The lower flows are mainly represented by massive varieties, and amygdaloid zones are widespread in the upper flows. In addition to lavas, there are also sills of crystalline gabbro-dolerite. The sporadic occurrence of pillow lavas suggests that sometimes eruptions occurred below sea level. North of the Pechenga area, there are numerous dykes striking northeast and similar in composition to volcanics of the Majärvi Formation. Therefore, one can suggest that the area of subsequent volcanism was wider than that of the initial sedimentation.

Basaltic andesites of the Majärvi Formation are distinctly enriched in light rare earth elements (REE), and show a negative europeum anomaly (Fig. 3). Two types of the chondrite-normalized heavy REE distribution patterns can be distinguished: one is normal or flat, and the other is oblique. These peculiarities in the REE distribution are related to the different degrees of sialic crust contamination within the basalt melt (Smol'kin *et al.*, 1995).

The Rb–Sr age of the Majärvi volcanics is 2324 ± 28 Ma (Balashov *et al.*, 1993b; Smol'kin *et al.*, 1995), and the comagmatic dykes in the Keiv structure (the Acha River) are about 2314 Ma old according to Sm–Nd dating (Fedotov and Amelin, 1992). Unfortunately, the study of zircons to date the Majärvi volcanics was unsuccessful.

The Lower Jatulian. The deposits correlative with the Lower Jatulian are united into the Kuetsjärvi Group, which comprises three formations. The first of them, the 400-m-thick Kuvernerinjoki Formation, rests upon the weathering surface of the basaltic andesite rocks of the Majärvi Formation. Its most complete sequences are exposed along the Kuvernerinjoki Creek and downstream of the Luottnjoki Creek (Zagorodnyi *et al.*, 1964; Predovskii *et al.*, 1974). The formation was also penetrated by Borehohe IX (Fig. 2). Its basal layer is composed of chlorite-carbonate schists 0.4 m thick. According to Negrutsa's data (cited from Smol'kin *et al.*, 1995), three rhythms can be distinguished in the formation succession. The lower two rhythms represent the gradational-rhythmic alternation of the following rocks: feldspar-quartz and arkose hematite-bearing psammites of variable grain size; less common gravelstones and hematite-martite quartzose conglomerates and siltstones; carbonate-rich claystones; and dolomites. The upper rhythm is composed of psammites with variable grain size and gravelstones in its lower part, and of dolomites and their breccia in the upper part. The lower part of the rhythm sequence exhibits thin horizontal layering, whereas its upper part shows the alternation of cross, wavy, and cross-wavy bedding, as well as characteristic structural elements, such as erosional surfaces, soft sediment deformations, and neptunian dykes. In the dolomite quarry (the Luottnjoki Creek), two horizons of stromatolitic dolomites separated by oncolitic dolomites were recovered. Stromatolites are represented by *Murmania sidorenkia* Lubts., which is similar to *Sundia* Butin from the Jatulian deposits of Karelia (Lyubtsov, 1979). According to the data of A.M. Akhmedov (cited from Predovskii *et al.*, 1974) and Negrutsa (cited from Smol'kin *et al.*, 1995), these deposits accumulated in a transgressive basin with weak bottom hydrodynamics and moderate seismicity. In their structure and composition, the metasediments

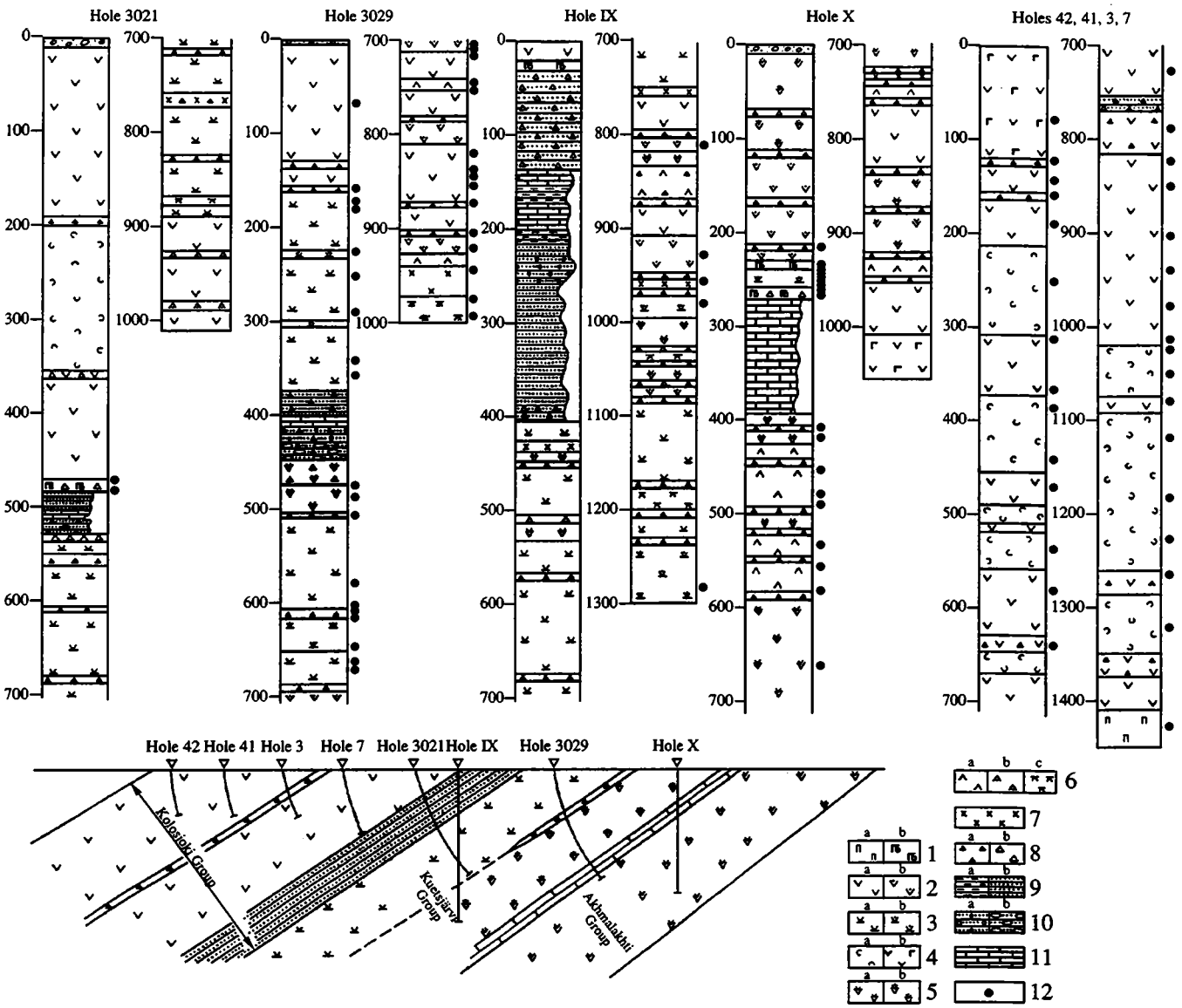

Fig. 2. Sections of the Akhmalakhti and Kuetsjärvi groups compiled on drilling materials (western Pechenga area).
1, (a) picrites and (b) picritic basalts; 2, (a) basalts and (b) subalkaline basalts; 3, (a) ferrobasalts and (b) subalkaline ferrobasalts; 4, (a) basaltic pillow lavas and (b) gabbro-dolerite sills; 5, (a) basaltic andesites and (b) subalkaline basaltic andesites; 6, (a) andesites, (b) trachyandesites, and (c) ferroandesites; 7, dacites and rhyolite-dacites; 8, (a) lava breccia and erruptive breccia with lava matrix and (b) tuffs; 9, (a) tuffites and clayey-silty and (b) silty-sandy metasediments; 10, (a) gravelstones and (b) conglomerates; 11, carbonate rocks; 12, location of samples.

of the considered formation are the most similar to deposits of the Lower Jatulian Yangozero Horizon of central Karelia, differing from the latter in a higher content of carbonate material (*Geologiya ...*, 1987).

Metasedimentary rocks of the Kuvernerinjoki Formation are conformably overlain by subalkaline volcanics of the Perttijärvi Formation, 150–250 to 1000 m thick. These volcanics exhibit a basic-to-acid trend: the lower part of the section is composed of tuffs and thin lava flows of picritic basalt, whereas its middle and upper parts consist of trachybasalt and basaltic trachyandesite. These latter form flows 0.5 to 25 m thick and horizons of lava breccia, tuffolava, and tuff. Lava flows often show an upper amygdaloid zone composed of more acid or alkaline (dacite and trachyte) rocks. Separate dacite sills of the subvolcanic type also occur. The Rb–Sr age of volcanics of the formation is 2214 ± 54 Ma (Balashov *et al.*, 1993c; Smol'kin *et al.*, 1995).

Subalkaline volcanic rocks are overlain by a well-traceable marker horizon from 10 to 40 m thick composed of conglomerates and shales. The horizon is often underlain and overlain by pillow lavas. The pebbles and lapilles are predominantly composed of underlying volcanic rocks, whereas the shales consist of vol-

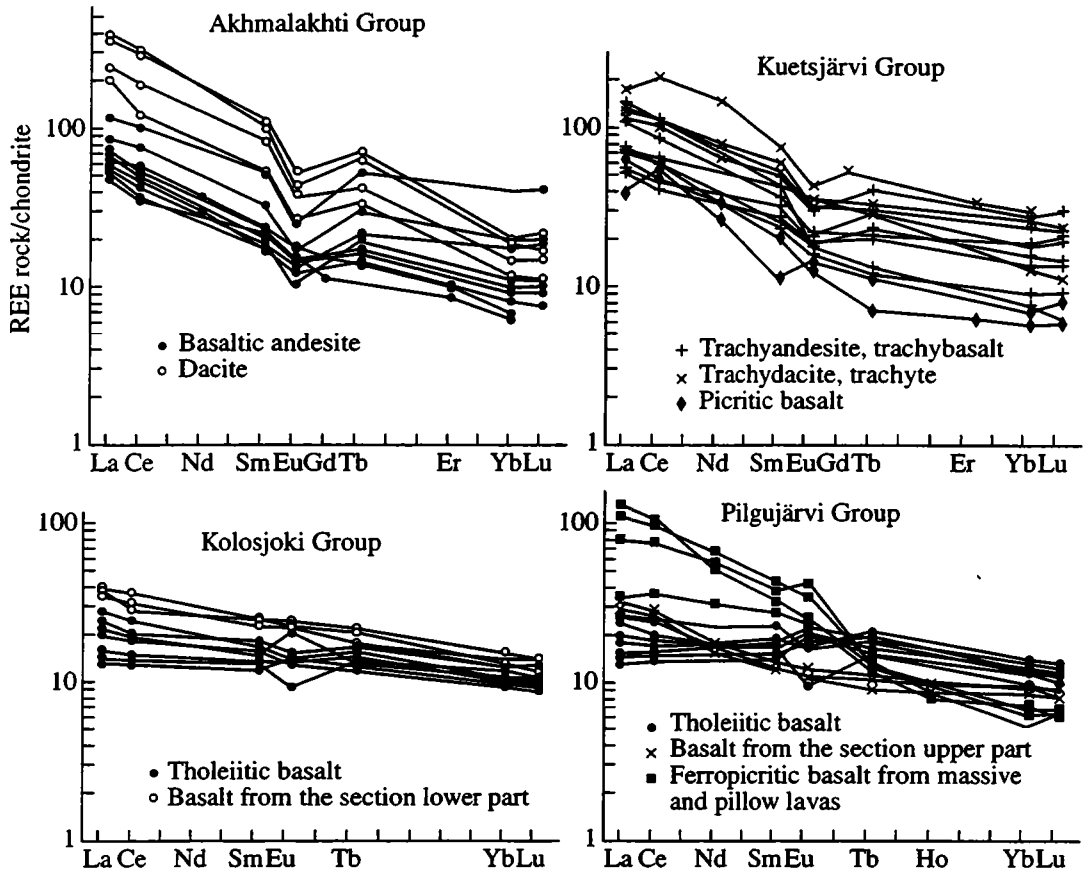

Fig. 3. Chondrite-normalized REE spectra for volcanics of the northern zone.
In addition to the original data, the data from Kremenetskii and Ovchinnikova (1986) on KSH-3 and Mints (1993) were used.

caniclastic material formed in marine environments. The most complete sequence of the horizon can be observed on the Orshoaivi Mt. eastern slope (Predovskii *et al.*, 1974) from where it extends in a latitudinal direction wedging out only in the western and eastern flanks of the Pechenga structure (Fig. 1).

Volcanics of mixed composition occur upsection. In contrast to earlier viewpoints, we consider these rocks to represent the separate Orshoaivi Formation, because they are distinct in the sections and laterally well-traceable. The most representative successions of the formation, which can be considered as its stratotype, is exposed on the Orshoaivi Mt. slopes. In addition, its complete sections are penetrated by the IX and 3021 boreholes (Fig. 2). The Orshoaivi Formation mostly consists of massive and, less commonly, pillow basaltic lavas with tuff interbeds and subvolcanic bodies of basaltic trachyandesite and dacite. Thus, the Pittijärvi and Orshoajvi formations distinctly differ in composition: the former is dominated by subalkaline varieties, and the latter, by normal-type volcanics. The alternation of volcanics of different types indicates the gradual replacement of magma sources. It should be noted that volcanics of similar type are almost absent in the Proterozoic deposits of Karelia and Finland, which testifies to the peculiar geotectonic regime responsible for the formation of the Pechenga structure and Pechenga-Varzuga belt as a whole.

The volcanic rocks of both formations are enriched in light REE, and sometimes exhibit the europeum anomaly (Fig. 3). The highest light REE contents are recorded in trachytes, whereas picritic basalts show the lowest concentration, which correlates well with the lower alkali content in the latter rocks. Moreover, the rocks of different composition show a gradual transition between each other, which may indicate their genetic relations. Volcanics of the basaltic andesite formation differ from those of the trachybasalt formation in their distinctly lower europeum anomaly and more flat distribution patterns of heavy REE.

All researchers share the opinion that the volcanics of the Kuetsjärvi Formation erupted in subaerial environments. This is confirmed by the presence of weathering crusts and black scoria with a high degree of Fe oxidization (Zagorodnyi *et al.*, 1964; Predovskii *et al.*, 1974, 1987). Basaltic eruptions were of the fissure type; occasionally they occurred under water, and produced pillow lavas. Eruptions of acid and alkaline melts were typical of the central-type volcanoes. One of these volcanoes is exposed by erosion near the monument to

the defenders of the Arctic in the city of Zapolyarnyi. Such volcanos produced in abundance the eruption breccia, containing fragments of granites, granite-gneisses, and amphibolites. In terms of the sequence structure, type of eruptions, and geochemical composition of volcanics, the Pirttijärvi Formation correlates well with the Umbino Formation in the central part of the Imandra-Varzuga area (*Problemy* ..., 1976; Predovskii *et al.*, 1987; Smol'kin, 1992).

The Upper Jatulian. The age analogs of the Upper Jatulian in the Northern Pechenga zone are assigned to the Kolosjoki Group, which rests upon the weathering surface of the Orshoajvi Formation volcanics. The basal part of the group is composed of red arkose sandstone, gravelstone, conglomerate, dolomite, jasper, and tuffite of the Luchlompolojoki Formation, with total thicknesses from 80 to 150 m and, sometimes, to 400 m. Its most complete section occurs north of the Nikel' Settlement, in the Luchlompolo Creek area (Zagorodnyi *et al.*, 1964; Polyak, 1968; Predovskii *et al.*, 1974). The formation was also penetrated by the IX and 3021 boreholes (Fig. 3). The gradual transition from coarse to fine-grained terrigenous and carbonate, and then to carbonaceous-siliceous deposits, along with parallel changes from massive to coarsely and wavy cross-bedded, and then to fine-wavy cross-bedded textures can be observed by going from the formation base upward. As it was noted by many researchers, the rocks are enriched in potassium which, in addition to their red color, is a characteristic feature of the formation. The underlying volcanics of the Pirttijärvi and Orshoaivi formations, along with the subvolcanic bodies intruding them, were the main source of the clastic material. The upper part of the red dolomites incorporates bioherms of stromatolite *Conophyton minusculum* Semikh., analogous to those described from the Aphebian deposits of Canada (Lubtsov, 1979). The formation sections terminate with the carbonate-chlorite and carbonate-sericite-chlorite schists 5–60 m in thickness, which were formed after the tuffites and vitroclastic tuffs of basaltic and/or picritic composition, as follows from their relic textures. In composition and section structure, the Luchlompolojoki Formation is closest to the lower part of the Upper Jatulian Tulomozero Horizon in central and northern Karelia (*Problemy* ..., 1976; *Geologiya* ..., 1987). The Upper Jatulian of Karelia displays the maximum abundance and diversity of stromatolite associations and intense carbonate accumulation, which are the result of the wide Jatulian transgression, which covered for the first time the Karelian craton (*Geologiya* ..., 1987; Semikhatov and Raaben, 1994). Transgressive sedimentation patterns in the Late Jatulian are also thought to be characteristic of the Pechenga basin (Smol'kin *et al.*, 1995).

Karhu (1993) showed that carbonate rocks of the Kuvernerinjoki and Luchlompolojoki formations distinctly differ in their carbon isotope composition (respectively, $\delta^{13}C$ is 7.0 ± 1.0 and $1.2 \pm 0.8‰$ on average). Similar C-isotope anomalies were also established by this researcher in other structures of Karelia and Finland, which testifies to the drastic change at the Early–Late Jatulian boundary.

The contact between the metasediments of the Luchlomplojoki Formation and volcanics of the overlying Zapolyarnyi Formation is sometimes disturbed (Luchlompolojoki Creek, KSH-3) by tectonic dislocations of different age. The base of the Zapolyarnyi Formation in the sections, which is located 2 km west of Luchlompolojoki Lake and penetrated by KSH-3 (depth interval 4616–4647 m), includes intensely foliated ferropicrite lava flows with an MgO content of 15–19% (Predovskii *et al.*, 1974; Kozlovskii *et al.*, 1984). They recorded the first occurrence of ultramafic volcanism in the Pechenga area (Smol'kin, 1992). Well-exposed sections of the formation can be observed in the Zapolyarnyi area and along the new Zapolyarnyi–Nikel' road between Zapolyarnyi and Pakhta.

In the central part of the Pechenga structure, the Zapolyarnyi Formation consists of two members separated by a discrete, relatively thin (up to 70 m) horizon of agglomerate and tuffaceous shale (Fig. 3). The total thickness of both members is 1600 m in natural outcrops, and 2000 m in the KSH-3 section. These members are composed of massive (72%) and less common pillow (23%) lavas, with separate layers of lava breccia, tuff breccia, tuff agglomerate, hyaloclastite, and tuff (5%); sills of ophitic gabbro also occur. Compared to the lower member, the flow thickness in the upper unit decreases, varying here from 0.5 to 20 m, and the degree of sequence contrast increases. The thickness of tuff is 3 to 5 m. Judging from the occurrence of pillow lavas and hyaloclastites, the volcanics of the Zapolyarnyi Formation were formed as a result of underwater fissure eruptions.

The chondrite-normalized REE spectra of volcanics of both members are similar to the MORB type, though they are somewhat enriched in light REE (Fig. 3). Therefore, they are referred to a single tholeiitic basalt formation, indicating an incipient stage of crustal extension. The Rb–Sr age of the volcanics is 2114 ± 52 Ma (Balashov *et al.*, 1993c; Smol'kin *et al.*, 1995). Based on the U–Pb zircons datings, the age of volcanics and dykes of similar composition from the Jatulian successions of central Finland (Piippolanmaki, Kitsu) is 2113 ± 4 Ma (Pekkarinen *et al.*, 1991).

The Ludicovian. The deposits referred to this superhorizon are subdivided into two tuff-sedimentary (Zhdanov and Lammas) and two volcanogenic (Matert and Suppvaara) formations composing the Pilgujäarvi Group (table). The Zhdanov Formation, which is mainly composed of terrigenous rocks, rests upon the basalts of the Zapolyarnyi Formation without visible signs of hiatus and angular unconformity. It incorporates numerous gabbro-wherlite intrusions containing industrial deposits of copper-nickel sulfide, and was earlier known as the "productive member." Its most complete section occur in the eastern ore field

(Zhdanov, Western, and Southern quarries) of the Pechenga region. In the central part of the ore field, it is penetrated by KSH-3 (Kozlovskii *et al.*, 1984). The formation continues to the eastern flank of the Pechenga structure (the Kuchin Mt. area), and to its western flank up to the Nikel'–Rayakoski road. Near the Russian–Norwegian boundary, it wedges out and then appears again on the Norwegian territory, south of Skogfoss (Melezhik and Sturt, 1994).

The overlying Lammas Formation is of variable thickness (from 5–10 to 400–600 m), and consists predominantly of basaltic and ferropicritic tuffs. Its lower boundary is defined by the appearance of psephitic tuffs intensely altered by hypergenic processes. The maximum thicknesses are recorded in the Lammas and Valas creeks area, and in the Kotselvaara quarry, whereas in the Kuchin Mt. area, the formation wedges out completely. In the Verkhnee Kierdzhipori, the prospecting boreholes penetrated several horizons of ferropicrite lava breccia with a total thickness of 600 m (Skuf'in and Fedotov, 1989); they mark the second level of ultramafic volcanism.

There are several subdivision schemes for the Zhdanov and Lammas formations. Uvad'ev (1968), and later, Predovskii *et al.* (1974), defined several horizons and rhythmically layered members in their successions. According to Bekasova (1978), these units correspond to a complete transgressive-regressive cycle divisible into three cycles of the second order. The first of the latter cycles consists of basal siltstonee, sandstonee, and sandy claystone members. The second cycle comprises phosphorite-bearing gravel-sandy claystone and carbonate claystone members. The third cycle is composed of tuff-sedimentary and volcanogenic members. The considered formations are characterized by the occurrence of black sulfide-carbonaceous shales with a high content of organic matter, lenses of phosphate-bearing conglomerates and siliceous shales, and carbonate, phosphate, siliceous, and sulfide concretions. In the Lammas Formation, lenticular pebble interbeds and scattered boulders of plagiogranite, granite gneiss, and jaspillite occur. In Gilyarova's (1976) opinion, these rocks can be interpreted as iceberg facies.

Lithologic studies show that during the Zhdanov and Lammas time intervals, the position of the sedimentation basins, changes in facies patterns and thicknesses of the sedimentary deposits were controlled by a system of synsedimentary faults, and high-magnitude block movements resulted in drastic changes in facies patterns even in the limited areas (Predovskii *et al.*, 1974; Bekasova, 1978). Three facies subzones can be distinguished: (1) the western subzone with the most deepwater deposits; (2) the central subzone, with predominant coarse-clastic deposits of fans and coastal shallow areas; and (3) the eastern subzone (Lammas Lake and Lammas Creek area), with a noticeably increased amount of tuff material (Golubev *et al.*, 1984). The role of basic tuff material is also significant in the coeval deposits of the Pasvik area in northern Norway (Melezhik and Sturt, 1994).

The deposits of the Zhdanov and Lammas formation were formed in a shallow basin under conditions of compensated subsidence and a stable tectonic regime, according to Predovskii *et al.* (1974, 1987) and Bekasova (1978). Unlike these authors, Negrutsa (in Smol'kin *et al.*, 1995) argues for the deepwater environments of sedimentation because of the turbidite rhythmicity of metasediments. Based on the section structure, rhythmicity and composition of the deposits, the Zhdanov and Lammas formations can be referred to the terrigenous-shaly flyschoid formation.

The central part of the northern zone is composed of volcanogenic deposits of basic and ultrabasic composition. In the Pechenga ore field, they are conformably overlain by deposits of the Lammas or Zhdanov formations, and in the western flank of the Pechenga structure, they unconformably rest upon the Pirttijärvi Formation rocks. The distribution area of the above-mentioned volcanics is divided by synsedimentary faults into the Kuorpukas, Matert, and Suppvaara blocks (Fig. 1). The first two blocks are composed of volcanics of the Matert Formation, and the third one comprises volcanic rocks of the Suppvaara Formation. The paleovolcanologic investigations and studies of materials from the new structural boreholes allowed us to subdivide the Matert Formation into several rock associations.

The lower of them with an average thickness of 1600 m, is composed of interlayering flows of massive and pillow basalts, horizons of basic lava breccia, tuff, tuffite and hyaloclastite, and lenses of sulfide-carbonaceous shales. The lower and upper parts of this sequence include layered picrite flows, which mark the third and fourth levels of ultramafic volcanism, respectively. The second rock association consists of siliceous tuff-sedimentary deposits representing a marker horizon traceable over a distance of more than 20 km from the Ostrovnoe Lake (in the east) to the Kuorpukas Mt., and further to the Russian–Norwegian boundary (in the west). With thicknesses from 5–10 to 200–300 m, it is composed of a regular alternation of fine-grained massive horizontally layered and cross-bedded chemogenic deposits of silicilith and tuff-silciilith types (often with carbonate and cherty concretions), and clastic gravel-sized deposits, which form several rhythms separated by erosional surfaces (Smol'kin, 1992). Cross-bedding and soft sediment deformations sporadically occur within the rhythms. Sometimes, the upper rhythm terminates with sulfide-carbonaceous shales. In the northern slope of the Kuorpukas Mt., the considered deposits rest upon ferropicrite lava flows, the top of which is marked by the montmorillonite crust that resulted from underwater weathering (Smol'kin, 1992). In addition to fragments of the underlying basalt, ferropicrite, jaspillite, and acid tuffs, the clasts include granophiric rapakivi-like granites similar to the

Archean rocks (2.7–2.6 Ga) from the area located north of the Inari Lake in the Lapland region of Finland.

As it is known, the high-silica rocks represent a typical member of the volcanic-tuff-siliceous formations of the Phanerozoic that were formed in the underwater volcanism areas. Judging from the wide distribution of pillow lavas and hyaloclastites, occurrence of halmyrolysis zones, and other features, eruptions of magmatic melts in the studied area also proceeded in underwater, probably, deepwater environments. The age of the rocks (1970 ± 5 Ma) was determined using the Pb–Pb method on zircon from coarse-clastic deposits in the Ostrovnoe Lake (eastern flank) area (Hanski *et al.*, 1990; Smol'kin, 1992).

The third rock association of the Matert Formation was penetrated by structural borehole 3077-IV (Shulgjaur Lake). It consists of massive and pillow basalt lavas, hyaloclastites, rare lenses of tuffs and sulfide-carbonaceous shales. The thickness of this unit is 1500 m. The middle part of the section (the Dvoinaya Mt. and Shulgjaur Lake localities) is composed of an alternation of massive and pillow lavas, and layered ferropicrite flows, which represent the fifth level of ultramafic volcanism.

The succession of the northern zone is crowned by the Suppvaara Formation, distinguished for the first time. It is distributed within the structure of the same name, which is bordered by faults and recognized in all magnetometric maps. The geological mapping showed that the volcanics of the Matert Formation wedge out toward the margins of the northern zone, whereas the Suppvaara volcanic rocks are traced to the southeastern margin of the southern zone, i.e., to the Tyulpvyd tundra and Peschanoe Lake areas (Fig. 1). Its type sections occur in the Malaya Pechenga middlestream, Suppvaara Mt., and Shulgjaur Lake areas. Unlike the Matert Formation, the Suppvaara Formation is of a uniform structure, and consists of monotonously alternated thick flows of massive and pillow basalts; its sequences lack ferropicrites and high-silica rocks. The thickness of the formation is up to 1700 m.

Volcanics of the Matert and Suppvaara formations are generally represented by massive (47% on average) and pillow (40%) basic lava, breccia (1%) and tuffs (8%). In addition, they are associated with ferropicrites and their tuffs (4%), with the share of pillow lavas increasing in the Suppvaara Formation up to 70%. Respective eruptions proceeded in underwater environments in a slope area with highly differentiated relief, which resulted in sharp variations of the flow thicknesses, facies transition from massive to pillow lavas, and the formation of turbidite horizons with gradational bedding. The greater thicknesses of basalt tuff in the Ostrovnoe Lake area and its gradual reduction toward the Kuorpukas block may indicate a southward migration of eruption centers. The volcanism was accompanied by the intrusion of dolerite and picrite-dolerite dykes of two generations, which cross sediments and nickel-bearing gabbro-wherlite intrusions of the Zhdanov Formation.

In the distribution area of the tuff-sedimentary and volcanogenic rocks of the Pilgujärvi Group, lenticular sills of ophitic gabbro are widespread. In terms of their geochemical features, they are close to tholeiitic basalts of the Matert Formation, and are therefore referred to their comagmatic varieties (Predovskii *et al.*, 1974; Kremenetskii and Ovchinnikov, 1986). Under tectonic deformations, they behave as monoliths and often occur in the frontal parts of thrusts, which complicates the formation structure.

The above-mentioned tholeiitic basalts and ophitic gabbro show flat REE distribution patterns that suggests their similarity to the MORB-type volcanics (Fig. 3). However, they differ from the latter in positive potassium, rubidium, and barium anomalies. This is probably a result of the contamination of parent magma with sialic material that is possible in the continental crustal setting. Tholeiitic basalts of the Zhdanov and Matert formations are almost identical in REE spectra. The volcanics of the Matert Formation from the central Pechenga structure and its eastern margin (the Kuchin Mt.) are also close to them in this parameter, despite the different degree of metamorphism (prehnite-pumpellyite and amphibolite facies, respectively). Metabasalts of the Suppvaara Formation in the northern and southern zones exhibit a noticeable enrichment in light REE. Ferropicritic volcanics and comagmatic intrusive gabbro-wherlites show strong enrichment in light REE and a somewhat lower content of heavy REE (Fig. 3); they are also enriched in FeO and Fe_2O_3 (more than 14%), Cr, Ni, Ti, Zr, P_2O_5, and S (Smol'kin, 1992; Hanski, 1992). Ferropicrites of the same facies types (massive and pillow lavas, layered flows, lava breccia, and tuffs) from different levels show significant geochemical similarity, which suggests a stable composition of their parent magma for a long period of time.

Geochemical and isotope data suggest the coexistence of two independent mantle sources: the depleted shallow magma source for tholeiitic basalts, and a metasomatically transformed deeper source for ferropicrites (Smol'kin, 1992). According to Balashov's data (in Smol'kin *et al.*, 1995), carbonates from calcareous concretions, limestones, and dolomites of the Kuvernerinjoki, Luchlompolojoki, and Zhdanov formations demonstrate a high initial $^{87}Sr/^{86}Sr$ ratio (0.7067–0.7095) which indicates a considerable influx of radiogenic strontium from the adjacent Archean rocks. Available geochemical and isotope data do not confirm the hypothesis assuming the existence of a spacious basin with oceanic crust during the Jatulian to Ludicovian, as it was suggested by Mints (1993) and supported by Melezhik and Sturt (1994).

The formation of layered ferropicrite flows was at 1980 ± 40 Ma, according to the Sm–Nd dating (mineral isochron); this value corresponds to the age estimates by the Re–Os and Pb–Pb methods, which are 1980 ± 67 Ma

and 1980–1995 Ma, respectively (Hanski *et al.*, 1990; Smol'kin, 1992; Smol'kin *et al.*, 1995). The Rb–Sr whole-rock age of the tholeiitic basalt is 1980 ± 34 Ma (Balashov *et al.*, 1993c; Smol'kin *et al.*, 1995). We consider these estimates to be of the reference rank for the Ludicovian Superhorizon in the Kola region. A similar Sm–Nd age (1980 ± 27 Ma) is characteristic of picrites from the Konchozero sill of the Suisaari Horizon in the Onega basin (Pukhtel' *et al.*, 1992), which substantiates the assignment of studied rocks to the Ludicovian. Tholeiitic basalts from the Koli-Kaltimo area of central Finland, whose age determined by the Sm–Nd and U–Pb methods is 1985 ± 80 and 1965 ± 10 Ma, respectively, were also formed at that time (Vuollo *et al.*, 1992). The Sm–Nd data evidences that the intrusion of dolerite dykes in the northern framing of the Pechenga structure took place somewhat later, i.e., 1908 ± 10 Ma ago (Fedotov and Amelin, 1992; Smol'kin *et al.*, 1995). The latter dating, evidently, indicates the minimal age of the Ludicovian rocks in the considered area. Enderbites from the Kolvitsa-Lapland granulite belt located south of the Pechenga structure were also formed at the same time, as their U–Pb age (1925 ± 80 Ma) corresponds with that of granulite metamorphism (Bibikova *et al.*, 1993a).

STRATIGRAPHY AND VOLCANISM OF THE KARELIAN COMPLEX IN THE SOUTHERN ZONE

The Por'itash fault, which is south of the peaks of the Por'itash and Lak-Por'itash mountains and, according to geophysical data, dips southward at angles of 50°–60°, is considered to represent the geological boundary between the northern and southern zones of the Pechenga structure.

The geological structure of the southern zone is controversial, because it includes different highly metamorphosed rocks deformed into isoclinal folds and experienced local dynamometamorphism. In the 1950s–1970s, three main interpretations of its structure were popular. At first, N.A. Kuryleva (1948, unpublished data) and later L.I. Uvad'ev, Yu.I. Filimonov, A.A. Predovskii, P.K. Skuf'in and other geologists united the rocks of the southern zone into the sedimentary-volcanogenic rock association (since the 1970s, the group), and considered it to be younger than the formations of the northern zone. Others (Ivanova, 1953, unpublished data; Zagorodnyi *et al.*, 1964) distinguished several rock associations in the southern zone, each of which was correlated to formations of the northern zone. A similar point of view was shared by Golubev *et al.* (1984), who considered the shales of the Kallojaur Lake area in the southern zone to be an analog of the "productive member" of the northern zone. Unlike these authors, Gilyarova (1976) and Polyak (1968) referred the rocks of the southern zone to the Tundra Group underlying the Karelian Complex.

The mapping and geological studies carried out in the southern zone resulted in the distinguishing of six formations and rock associations there, the lowest of which (the Suppvaara Formation) is traced from the northern zone and belongs to the lower Karelian Complex; the five other units are restricted in distribution by the southern zone only, and referred to the upper Karelian complex or the Kalevian (table).

The Kalevian. In the Russian territory of the Baltic Shield, the reference sections of the Kalevian (earlier termed the Livvian) are located in the Ladoga and Onega lakes area (*Geologiya ...*, 1987; Semikhatov *et al.*, 1991). They are represented by rocks of the Ladoga and Besovets groups, which mainly consist of terrigenous rocks: conglomerates, arkoses, quartzites, and various shales. These groups overlie, with angular unconformity, the Ludicovian and Jatulian rocks, and those of Archean basement. No complete analogs of these Kalevian deposits were recorded in both the studied areas of the Pechenga structure and entire Pechenga-Varzuga belt. Yet, in northern Finland (central Lapland), the Kalevian level is marked in addition to conglomerate, quartzite, and sandstone by the basic, acid, and subalkaline volcanics of the Lainio Group 1.88 Ga old, which are overlain by conglomerates and sandstones of the Kumpu Group. As it is shown below, a similar situation is also characteristic of the southern Pechenga zone, which undoubtedly indicates a different geotectonic regime within the Karelian and Kola cratons during the Kalevian time.

We subdivide the Kalevian deposits of the southern zone into two groups: the Ansemjoki and Porojärvi. The former group incorporates the Kallojavr, Bragino, and Mennel' formations, whereas the latter consists of the Kaplya and Kassesjoki rock associations. The tuff-sedimentary rocks of the Kallojaur Formation, which compose the lower part of the succession in the southern zone, directly rest upon the Suppvaara Formation (basalts in the central area of the structure) or the Matert Formation (in its western margin) (Fig. 4; table). Prospecting trenches through the base of the formation (northwest of the Porojärvi Lake) revealed the conformable contact of a thin member of andesite tuff, tuff sandstone, and tuff gravelstone with underlying nonfoliated pillow lavas of the Matert Formation. The contact surface dips southward, with an angle of 35°. Upsection, the tuffites are subsequently replaced by members of rhythmically layered carbonaceous shales, siliciliths and carbonate-bearing phillites (450 m), sericite and sericite-carbonate siltstones (700 m), tuff sandstones and phillites (200 m), and finally, carbonaceous sericite-carbonate metapelites and metasiltstones (450 m); separate basalt and picritic basalt flows also occur. In the Por'itash and Lak-Por'itash Mts. area, shales associate with extrusive dacite and rhyolite bodies surrounded by the contact-metamorphism zone. Besides sulfide-carbonaceous shales, the Kallojaur Formation contains highly carbonaceous shales with a low sulfur content, which may indicate the decreased

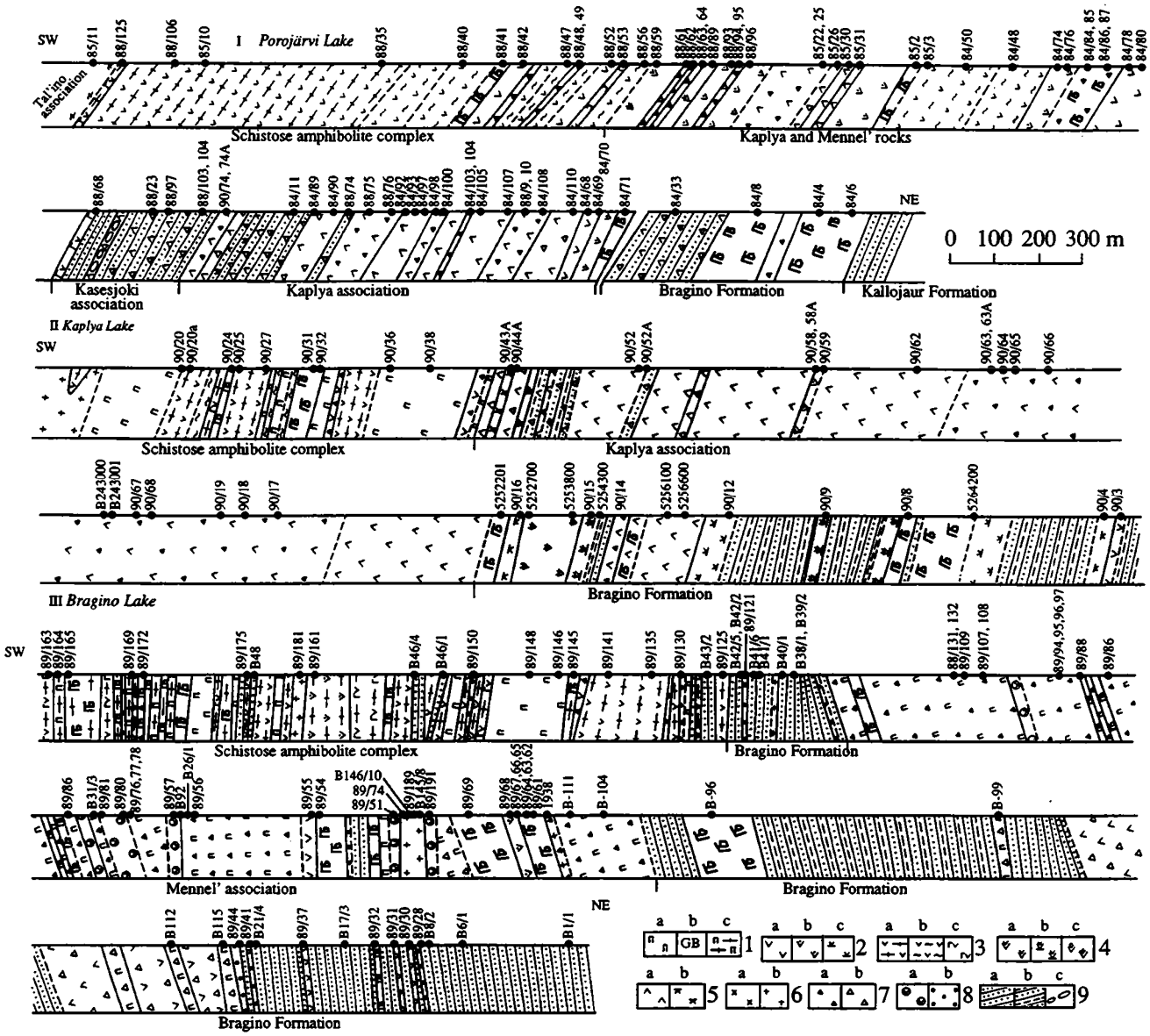

Fig. 4. Schematic sections in the Porojärvi, Kaplyajavr, and Bragino reference areas of the southern zone.
1, (a) picrites, (b) picritic basalts and magnesian basalts, (c) schistose amphibolites after picrite or picritic basalt; 2, (a) basalts, (b) subalkaline basalts, and (c) ferrobasalts; 3, (a) schistose amphibolites after basalt, (b) foliated basalts, and (c) gabbro-dolerite sills; 4, (a) basaltic andesites, (b) ferrobasaltic andesites, and (c) subalkaline basaltic andesites; 5, (a) andesites and (b) ferroandesites; 6, (a) dacites and (b) subalkaline rhyolites; 7, (a) lava breccia and eruptive breccia, (b) tuffs; 8, (a) alkaline basalts of mixed nature and (b) quartz-magnetite rocks; 9, (a) sandstones, (b) alternation of sandstones and claystones, and (c) tuff conglomerates.

influence of synchronous volcanism and exhalation processes, and an increase in basin depth. Nearly all of the Kallojaur sediments are consistently depleted in titanium, iron, and some rare elements; they usually show an admixture of tuff material in contrast to the metasediments of the Zhdanov and Lammas formations.

Observations by Akhmedov (in Golubev *et al.*, 1984) and Avedisyan (in Smol'kin *et al.*, 1995) show that sedimentary deposits of the Kallojavr Formation are characterized by a combination of flyschlike rhythmical, horizontal, and fine lamination, as well as slumping structures similar to those recorded in deep-water turbidites of modern oceans. In general, sedimentary rocks of the Kallojaur Formation differ from the Matert and Lammas sediments in the structural patterns of the transgressive-regressive cycles, which primarily resulted from the different basin depths, different rate of their subsidence and accumulation of terrigenous material, as well as from climatic changes.

A sequence of volcanics and sedimentary rocks is considered to represent the independent Bragino Formation. It has a two-member structure, and begins with thick massive and pillow basalt lava, lava breccia, hyaloclastite, and shale horizons, which are intruded by

ophitic metagabbro sills. This member has a maximum thickness in the Ansemjaur Lake area, and completely wedges out west of the Porojärvi Lake and east of the Pechenga River. The upper part of the unit consists of volcanics alternating with sediments and tuffs, which mark the transition to the overlying tuff-sedimentary rock association of the same formation. Proportions between different rocks in this member, whose thickness is 400–700 m at the flanks of the southern zone and 2500 m in its central areas, significantly varies along the strike: in the Porojävri Lake area, pyroclastic andesitic rocks dominate; in the Kaplya Lake area, an equal tuff/tuffite ratio is observed; and in the Bragino Lake area, the sequence mainly consists of sericite-bearing sedimentary rocks. Several types of schists are distinguished among tuff-sedimentary rocks of the Bragino Formation: (1) chlorite-carbonate and biotite-chlorite-carbonate (intermediate tuffs); (2) chlorite-carbonate and tremolite-chlorite-carbonate (picritic tuffs); (3) carbonaceous-carbonate, carbonaceous-chlorite, and sericite-chlorite (tuffites); and (4) carbonaceous and carbonaceous-sericite (sediments with a tuff material admixture). The upper unit of the formation incorporates numerous sills of granophyric basaltic andesite, basalt, and picritic basalt. Thus, the early stage in the formation history of the southern zone was already characterized by contrast volcanism, when explosive andesite eruptions were associated with the effusive activity of basaltic and picrite-basaltic magma.

The middle part of the succession in the southern zone is represented by volcanics of two different associations, which we assign to either the Mennel' or Kaplya units (rock associations). The direct age relationship between these two rock units remains unknown because they are separated by a fault and occur within a single narrow zone, which is also bounded by faults. The essential feature favoring the different age of these rocks is that the Mennel' rocks continue the geochemical trends established for the Bragino volcanics, whereas the Kaplya rocks show independent, strongly contrasting trends (see below). Based on these peculiarities, we consider these members not only to be diachronous, but also belonging to different, Ansemjoki and Porojärvi groups (table).

The most complete section of the Mennel' volcanics is exposed in the Bragino Lake area, where their thickness is up to 2000 m. The main part of the sequence consists of picrites, picritic basalts, and basalts; subvolcanic rhyolite-dacite and subalkaline rhyolite rocks are strongly subordinate picrites (60% of the sequence volume) and mainly occur as lava breccia or clastic lavas and, less commonly, as tuffs. Picritic and normal basalts form massive, amygdaloid or variolite, and, less commonly, pillow lava flows sometimes separated by hyaloclastite, tuff, and autobreccia horizons. Judging from observations of pillow lava flows, the thickness of the member was growing in a southward direction.

Separate fragments of the Mennel' rock association were observed along the western coast of the Porojärvi Lake in Russia, and near the Pasvik Lake in Norway. In these areas, this association, highly variable in composition, consists of clastic and tuffaceous lavas of picritic composition, rare massive to pillow lava flows, horizons of coarse-grained tuffs, and volcanomictic sandstones, siltstones, and claystones composed of picrite material. Sedimentary strata show the rhythmic bedding and synsedimentary slumping structures; they also include neptunian dykes.

Southwest of the Bragino and Mennel' lakes (west of the Kaskel' dome), there is a member of schistose amphibolites, which was earlier correlated to basalts of the Pilgujärvi Formation (Zagorodnyi *et al.*, 1964; Predovskii *et al.*, 1987). Our studies showed that these amphibolites were formed after picrites, picritic basalts, and basalts whose geochemical characteristics, the REE spectra included (Fig. 5), were similar to those of the Mennel' volcanics. This allows the schistose amphibolites be considered as fragments of the latter unit, but not as a part of the Matert Formation of the northern zone.

Two types of the chondrite-normalized REE spectra are distinguished in the Bragino Formation rocks (Fig. 5): (a) the flat spectrum similar to that in the tholeiitic basalts of the Matert Formation; and (b) the spectrum somewhat enriched in light REE. The second spectrum type may be a result of basalt melts interaction with the granite-gneiss basement, which is confirmed by geochemical data (Smol'kin *et al.*, 1995). Thus, a distinct change in the composition of volcanics and regime of magma melt ascent and eruption occurred during the Bragino time.

The rocks of both the Mennel' unit and schistose amphibolite block are characterized by enrichment in light REE. Despite their different composition (picritic and basaltic, respectively), these rocks show gradual transitions between each other that suggests their genetic relation. The volcanics are dissected by granite-syenite-porphyre veins, which are strongly enriched in light REE and depleted in heavy REE. Sometimes, the vein rocks show a negative europium anomaly.

As it was noted, the younger rocks of the southern zone are thought to compose the separate Porojärvi Group, which comprises the volcanogenic Kaplya and terrigenous Kasesjoki units. The most complete section (2000 m thick) of the Kaplya rock association is exposed in the eastern coast of the Porojärvi Lake. There, the section is composed of tuff lavas with rhyolite-dacite clasts and the andesite matrix (50%), agglomerate tuffs (10%), tuff sandstones and gravelstones (30%), and extrusive andesites (10%). Southeast of the Porojärvi Lake, the most common are basalts, magnesian andesites, and basaltic andesites (up to 70%); subordinate rocks are represented by tuff lavas and andesitic basalt tuffs. The second most complete section is in the Kaplya Lake area, where it mainly con-

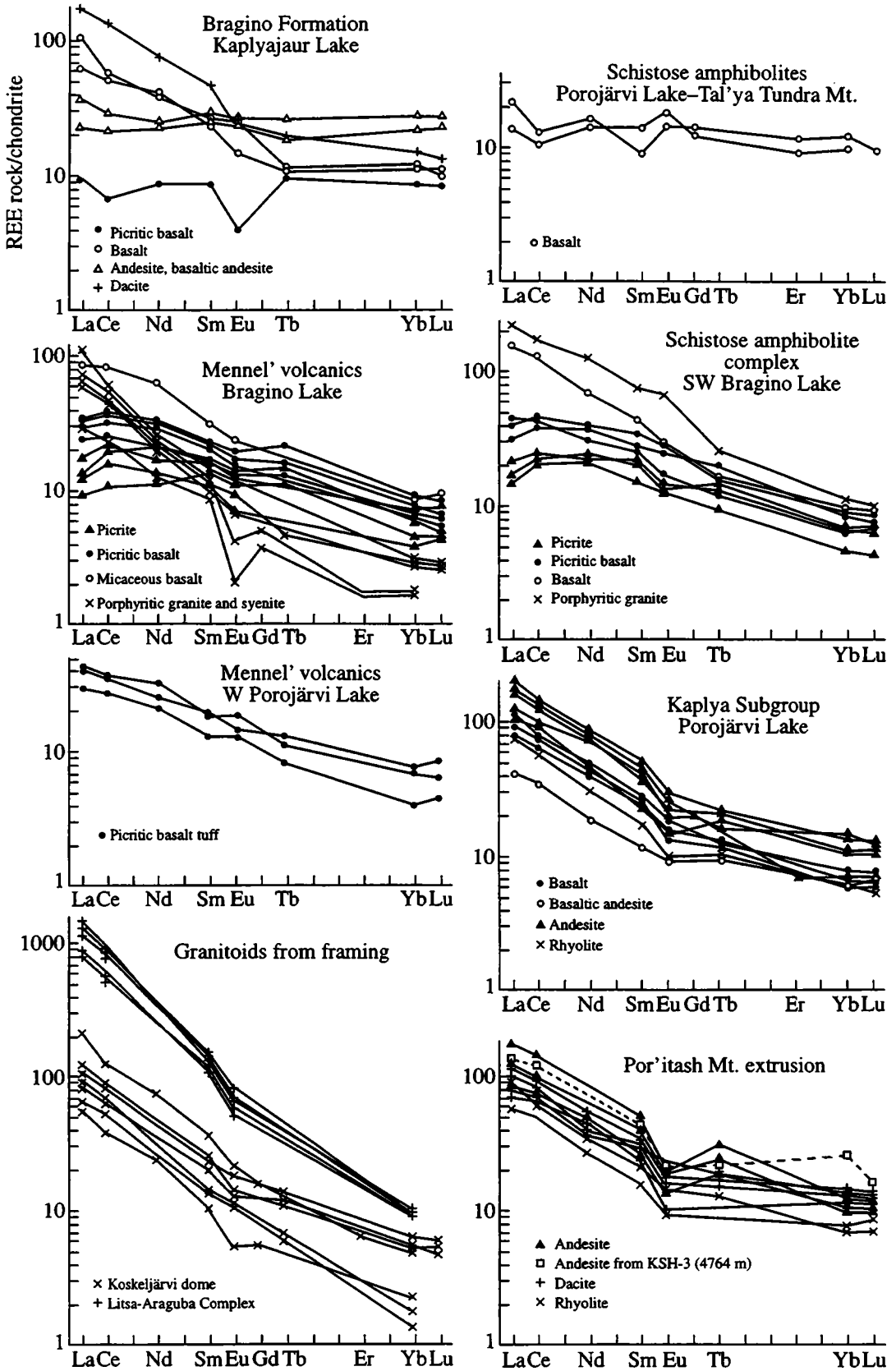

Fig. 5. Chondrite-normalized REE spectra for volcanics of the southern zone and granitoids of its framing. In addition to the original data, the data from Mints (1993) for granitoids were used.

sists of extrusive andesites and andesite-dacite tuff lavas (80%), with a subordinate amount of andesite lavas and tuffs.

The Kaplya volcanics and extrusive rocks in the Por'itash Mt. area show relative enrichment in light REE. The principal difference between them is evident from a negative europium anomaly consistently recorded in extrusive rocks that can be related to the feldspar fractionation. Judging from the REE patterns, the acid volcanics and extrusive rocks belong to a single genetic series. Still, in their REE spectra and concentrations of rock-forming components, both rock types distinctly differ from the granite-granodiorites of the Shuoni and Kaskel' domes, whose Pb–Pb age is 2180–2140 Ma (Mints *et al.*, 1989) and 1945–1940 Ma (Vetrin, 1988), respectively. These differences prevent the uniting of the rocks of extrusions and domes into a single comagmatic complex, as was previously suggested by Vetrin (1988) and Mints (1993).

Metasedimentary rocks of the Kasesjoki association compose the block which is situated east of the Porojärvi Lake and bounded by sublatitudinal faults (Fig. 1). The maximum thickness of these rocks (400 m) is recorded in the central part of the block; southeastward and northwestward, it sharply decreases. At the base of the association exposed in the northern flank of the block (the Kasesjoki Creek area), there is the 40-m-thick member of intensely foliated tuff conglomerates. These rocks show signs of gradational structures: the basal part of every microrhythm is composed of tuff agglomerates and tuff gravelstones, gradually replaced upsection by psephitic and psammitic tuff sandstones. The clasts are represented by volcanics of variable composition, including granophyric andesites and rhyolites with fluid texture. The matrix corresponds in composition to the subalkaline basalt. Rhythmically layered graywacke, sandstones and siltstones occur higher in the section. In the distribution area of the association, ferrodolerite sills (from 5 to 15 m thick) are widespread. In its lithologic features and composition of sedimentary and subvolcanic rocks, the Kasesjoki association differs from all of the studied formations and other rock units. One can suggest that it is younger than all of them, and terminates the rock succession in the southern zone.

Currently, there is no reliable geochronological information on the rocks of the southern zone. The whole-rock Rb–Sr datings for samples of basalt (the Ansemjaur Lake area), picrite (the Bragino Lake area), and andesite-rhyolite (the Porojärvi Lake area) range from 1865 ± 58 Ma to 1855 ± 54 Ma (Balashov *et al.*, 1993c), and evidently mark only the minimal age limit of the studied rocks. The intermediate and acid volcanics of the Latvajärvi Formation (the Lainio Group) in central Lapland, Finland are probably coeval; they were dated by the U–Pb method as 1883 ± 5 Ma old (Lehtonen *et al.*, 1992).

One of the key points of southern zone geology concerns the stratigraphic position of the extrusive massifs in the Por'itash and Lak-Por'itash Mts. area. These massifs, forming a series of en-echelon arranged lenticular bodies, intrude deposits of the Kallojaur Formation and contain xenoliths of the latter. Similar rocks are penetrated by KSH-3 at the depth of 4673–4684 m (Kozlovskii *et al.*, 1984; Kremenetskii and Ovchinnikov, 1986). The central and top parts of the thickest intrusions are composed of leucocratic dacites, whereas their andesite-dacite from basal parts are enriched in pyroxene phenocrysts. The andesite-dacites are intruded by lens-like and nearly isometric bodies of rhyolite breccia representing the second phase eruptions. In the REE spectra, the rocks from extrusive massifs are very similar to the Kaplya sedimentary rocks, and differ from granitoids of the Litsa-Araguba orogenic complex (Fig. 5). The anomalously high $^{87}Sr/^{86}Sr$ ratio (0.7088 ± 19) in dacites and rhyolites, and occurence of several zircon morphotypes in these rocks including those with 2520 and 2440 Ma datings by the ion thermoemission method (Balashov *et al.*, 1993c; Smol'kin *et al.*, 1995), suggest a significant contribution of crustal material in the magmatic melt generation. The Rb–Sr datings of the whole-rock samples (1718 ± 29 Ma) coincide with the age of regional metamorphism, and therefore, they cannot indicate the time of formation of aforementioned massifs. These facts suggest the comagmatic nature and synchronism of volcanics of the Kaplya rock association and subvolcanic bodies exposed on Mt. Por'itash.

Based on the general structure of the southern zone sequences, composition peculiarities of tuff-sedimentary and volcanic rocks, and analysis of their geochemical features, we distinguished among them the tuff schist, picrite-basalt-andesite, andesite-dacite-rhyolite, and conglomerate-schist formations and the dacite-rhyolite subvolcanic complex (table), all mainly characteristic of compressive environments.

All the rocks of the southern and northern zones experienced diachronous metamorphism, which resulted in the formation of single regional metamorphic zonality. According to data on Sm–Nd, Re–Os, and Rb–Sr systems in magmatic rocks and sulfide ores, the peak of metamorphism occurred at 1750–1700 Ma (Smol'kin *et al.*, 1995), simultaneously with granitization in the Archean basement. For instance, the U–Pb age of zircon from granite veins in the basement, which were penetrated by KSH-3, is 1754 ± 7 Ma (Mitrofanov *et al.*, 1991).

It should be noted that the previous analysis of magmatic rocks in the entire Pechenga-Varzuga belt, which is referred to the Early Proterozoic riftogenic system, allowed four stages in its geotectonic evolution to be defined: pre-riftogenic, early riftogenic, late riftogenic, and orogenic (Smol'kin, 1992, 1993). The pre-riftogenic stage (2.5–2.3 Ga) gave rise to mainly mantle-crust rock associations with normal alkalinity and a high initial

$^{87}Sr/^{86}Sr$ ratio (0.7042), and to layered peridotite-pyroxenite-gabbronorite intrusions with negative values of ε_{Nd} (–1, –2). During the early riftogenic stage (2.3–2.2 Ga), subaerial eruptions of mantle melts generally produced rocks with elevated alkalinity values and $^{87}Sr/^{86}Sr = 0.7035$. The late riftogenic stage, or the stage of maximum extension in the riftogenic system, is distinguished by the decreasing thickness of sialic crust, formation of deepwater troughs, and intrusion of melts from two independent sources, the shallow and deep, respectively, responsible for tholeiitic basalt ($^{87}Sr/^{86}Sr = 0.7021$) and ferropicrite ($^{87}Sr/^{86}Sr = 0.7032$; $\varepsilon_{Nd} = +1.5$; $^{187}Os/^{186}Os = 0.935 \pm 0.031$) eruptions. At the end of the stage, partial mixing of these two sources took place in intermediate chambers. During the terminal or orogenic stage (1.9–1.8 Ga), mantle melts ($^{87}Sr/^{86}Sr = 0.7029$) were replaced by mantle–crustal and crustal melts ($^{87}Sr/^{86}Sr = 0.7043$), which was followed by orogenesis and zonal metamorphism (1.8–1.7 Ga). Thus, in case of the Pechenga structure, we are dealing with a complete cycle of the riftogenic system development, from its initiation to complete opening, and then to closing and orogenesis.

CONCLUSION

(1) The Pechenga structure, a part of the Early Proterozoic Pechenga-Varzuga belt with a long (more than 500 Ma) evolution history, is one of the few examples of riftogenic systems which exhibit the complete evolutionary cycle from initiation and rifting (though without formation of oceanic crust) to closing and orogenesis. The considered area comprises two structural zones (northern and southern), both representing a single stratigraphic succession of the Karelian (Lower Proterozoic) Complex. The latter rests with angular unconformity on deeply metamorphosed rocks of the Late Archean age (2.9–2.7 Ga) and layered intrusions of the Early Proterozoic (2.50–2.45 Ga).

(2) The Karelian Complex of the monoclinal northern zone is more than 8.5 km in total thickness. It was subdivided into 11 tuff-sedimentary and volcanogenic formations, which can be referred to the Sariolian, Lower and Upper Jatulian, and Ludicovian superhorizons. Rocks of the intensely dislocated southern zone are divisible into five formations and rock associations of mixed volcanogenic and tuff-sedimentary composition. These rocks, up to 3 km in thickness, are assigned to the Kalevian Superhorizon, which terminates the Karelian Complex sequence in the studied region.

(3) Development of the Pechenga paleobasin was initiated on the Archean continental granite-gneiss crust after endogenic activization, the formation of layered intrusions, and long-term peneplenization. During the paleobasin evolution, the basal conglomerate formation (the Sariolian) was subsequently replaced by quartzite-carbonate, terigenous-tuff-carbonate, and terrigenous (shaly) or flyschoid formations (the Jatulian-Ludicovian). At the stage of paleobasin closing (the Kalevian), the conglomerate-shale or molassoid formation was deposited.

(4) Volcanogenic associations of the Pechenga structure were formed during six megacycles, which were separated by periods of dominating sedimentation, and differed in composition and environments of volcanic activity. The succession shows the transition from mantle-crust (the Sariolian) to depth-differentiated mantle (the Ludicovian), and then back to mantle-crust (the Kalevian) rocks. Active volcanism in the southern zone (1865–1855 Ma) was close in time to granite formation in the Kolvitsa-Lapland granulite belt (1884 ± 18 Ma), and to the Svecofennian activization of mantle and crust processes in Karelia and Finland (1900–1860 Ma).

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Conodonts from the Yarovka (Lower Silurian) Horizon of the Altai Mountains

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INTRODUCTION

This study is devoted to the Early Silurian conodonts, which were distinguished in an upper portion of the Yarovka Horizon stratotype (Inya River) and in the coeval deposits exposed along the Gromotukha River (Fig. 1), in the northwestern Altai Mountains.

The Llandoverian terrigenous-carbonate deposits, which are exposed on slopes and at the base of Mt. Rossypnaya, the left bank of the Yarovka River, were originally described as the Yarovka Beds (Kul'kov, 1967) and later suggested to be the stratotype of the respective horizon (Kul'kov, 1974). The stratotype section contains fauna of Tabulata, Rugosa, and Brachiopoda. Near the top, it has an interlayer of graptolite shale. The late Llandoverian age of this interval is inferred from the graptolite assemblage of *Retiolites angustoides* (Elles et Wood), *Monoclimacis griestonensis kettneri* (Boucek), *Monograptus priodon* (Bronn), and *Oktavites falx* (Suess) (Sennikov, 1976). The assemblage associates with brachiopods, including species of *Brachiprion arenacea* (Dav.) (Kul'kov, 1989). However, many of the brachiopods either belong to endemic species of limited significance for age determinations, or they are identified only at the generic level because of their poor preservation. The graptolites, in turn, are rare, and they occur at one or two levels near the top of the section. In addition, deposits of the Yarovka Horizon are represented by a series of rocks of different sedimentary facies that frequently replace one another in a lateral direction; this is well-evident from the relations between sections I and II, which are described below. As a consequence, the faunal communities of the Yarovka time changed their composition depending on different ecological environments with respectively different sedimentary facies, which presents additional problems for correlating the sections under consideration. Conodonts, which occur in both the shell-bearing and graptolite facies, seem to be a useful group for refining the age of the Yarovka Horizon. So far, the Silurian conodonts have not been distinguished in the Altai, and this study in some way settles the score.

MATERIALS

I sampled two sections in the stratotype locality: on the south-southeastern (I) and eastern (II) slopes of Mt. Rossypnaya. Two other sections (III and IV) were sampled on the left-bank outcrops of the Gromotukha River. I am very grateful to M.Yu. Drozhzhin and T.V. Drozhzhina, who actively assisted in this work. The dissolution of samples was performed by S.V. Grishin, and D.A. Andreev kindly assisted me in photographing the conodont species.

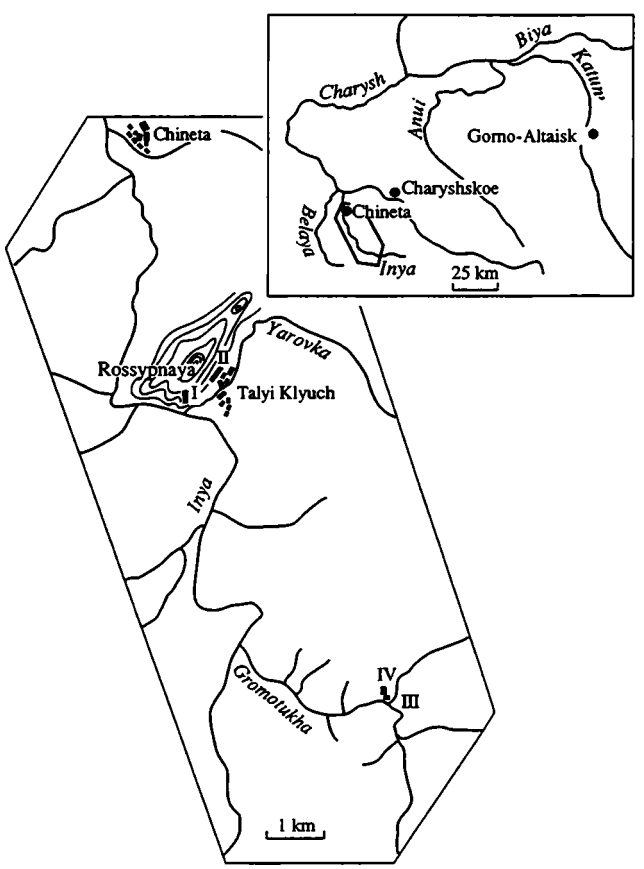


Fig. 1. Location scheme of studied sections (marked by numbers I–IV).

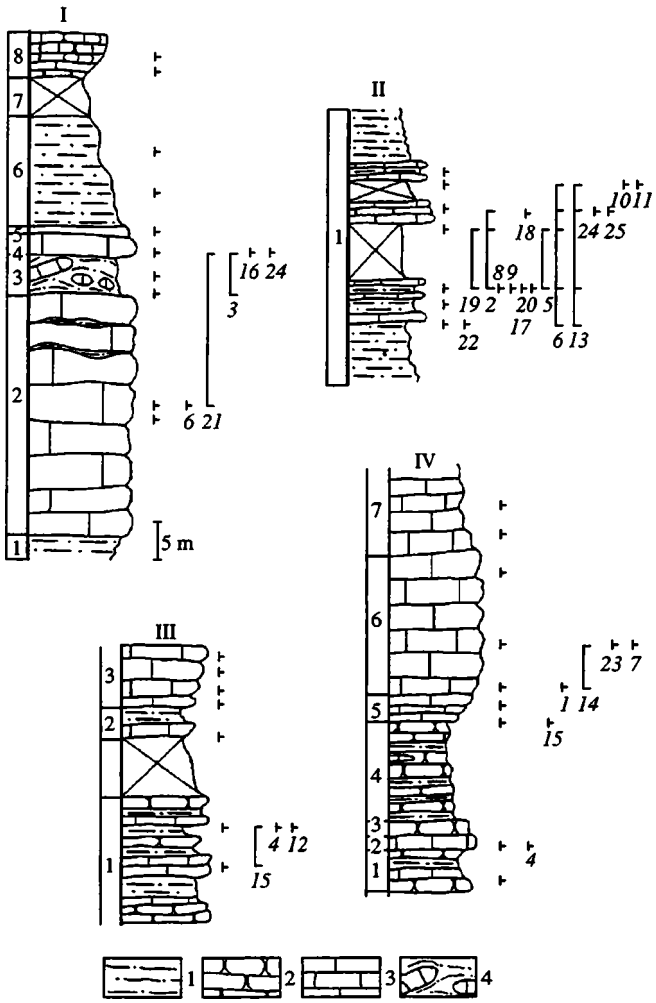


Fig. 2. Positions of conodont species in studied sections I–IV (with bed numbers 1, 2, 3, etc.).

1, siltstones and graptolite-bearing shales; 2, sandstones; 3, limestones; 4, mixtite horizon.

Conodont species (by numbers in italic): *1*, *Apsidognathus tuberculatus* s.l.; *2*, *Ap. t. arcticus*; *3*, *Ap.* n. sp. Armstrong 1990; *4*, *Ap.* sp.; *5*, *Aulacognathus bullatus; 6*, *Distomodus staurognathoides*; *7*, *D.* sp.; *8*, *Johnognathus huddlej*; *9*, ? *Oulodus fluegeli*; *10*, *Oulodus* sp.; *11*, *Ozarcodina* cf. *excavata*; *12*, *Ozarcodina pirata*; *13*, *Panderodus gracilis*; *14*, *P. unicostatus*; *15*, *P.* sp.; *16*, *Pseudobelodella silurica*; *17*, *Pseudoneothodus tricornis*; *18*, *Pterospathodus amorphognathoides*; *19*, *P. celloni*; *20*, *P. pennatus pennatus*; *21*, *P.* cf. *praeamorphognathoides*; *22*, *P.* sp.; *23*, *Tuxekania* sp.; *24*, *25*, Gen. and sp. indet 1, 2.

Brief descriptions of the studied sections (Fig. 2) and determinations of the discovered conodonts are presented below.

Section I. On the southern slope of Mt. Rossypnaya, 150 m upstream from the mouth of the Yarovka River, the following units are exposed (from the base upward):

(1) Siltstone, greenish- and yellowish-gray, with interlayers of greenish-gray fine-grained sandstone and rare horizons (5–10 cm) of flattened clayey-calcareous segregations. Thickness: 50 m.

(2) Limestones, massive, organic-detrital, gray, blue-gray, and patchy pink-gray. In the upper third of the unit, the rocks contain twisting interlayers (2–10 cm) of green and red sandy siltstones and the conodonts *Distomodus staurognathoides* Walliser and *Pterospathodus* sp. Thickness: 30 m.

(3) Mixtite horizon with boulders (up to 40 cm in diameter) and fragments of organic-detrital limestone beds, which are gray to pink-gray in color and embedded in a matrix of alternating layers (10–20 cm) of red-brown and green siltstones. The limestones yielded conodont species of *Pterospathodus celloni* (Walliser). Thickness: 4.5 m.

(4) Limestone, gray, organic-detrital, with green and yellow interlayers rich in clayey-sandy material. The unit contains the conodonts *Apsidognathus tuberculatus* Walliser, *Pseudobelodella silurica* Armstrong, and *Pterospathodus* sp., Gen. and sp. indet 1. Thickness: 2 m.

(5) Siltstone, greenish-gray, with thin interlayers of fine-grained sandstone and rare horizons (5–10 cm) of flattened calcareous segregations. Thickness: 1 m.

(6) Black graptolite-bearing shale. Thickness: 13–15 m.

(7) Unexposed interval with scattered fragments of greenish-gray siltstone and isolated outcrops of porphyritic diabase dykes. Thickness: 5 m.

(8) Basal limestone of the Chagyr Horizon, barren of conodonts.

Section II on the eastern slope of Mt. Rossypnaya, the left bank of the Yarovka River opposite the Talyi Klyuch, exposes the middle part of a 120-m-thick sequence of siltstone deposits, which include the following bed:

1. Siltstones, greenish-yellow, yellow, or gray, with interlayers of gray to dark-gray crystalline limestones. The limestones contain rare fragments of brachiopod shells associated with abundant fauna of Tabulata, Rugosa, Crinoidea, and Bryozoa. Samples selected across the bed strike yielded the following conodont assemblages (from the base upward):

(1) *Distomodus staurognathoides*, *Panderodus gracilis* (Branson and Mehl.), and *Pterospathodus* sp.;

(2) *D. staurognathoides*, *Apsidognathus tuberculatus arcticus* Armstrong, *Johnognathus huddlej* Maschkova, *Pterospathodus celloni, P. pennatus pennatus* (Walliser), *Aulacognathus bullatus* (Nicoll and Rexroad), ? *Oulodus fluegeli* (Walliser), *Pseudoneothodus tricornis* (Drygant), and *Panderodus gracilis*;

(3) *D. staurognathoides* and *A. tuberculatus arcticus*;

(4) *D. staurognathoides, A. tuberculatus, Pterospathodus amorphognathoides* Walliser, *Aulacognathus bullatus, Ozarcodina pirata* Uyeno and Barnes, and *Panderodus gracilis*;

(5) *D. staurognathoides, Ozarcodina* cf. *excavata* Branson and Mehl., *Panderodus gracilis*, and *Oulodus* sp.

Section III. In the steep right bank of the Gromotukha lower course, 200 m above the mouth of an unnamed tributary, the following units are exposed (from the base upward):

(1) Alternating beds (1–5 cm) of siltstone, greenish-yellow fine-grained sandstone, and gray sandy limestone. The sandstone beds contain interlayers (1–3 cm) that are enriched in carbonate material and show a cellular structure on the weathered surface. Two levels (7 and 11.5 m above the base) yielded specimens of *Panderodus* sp., and such forms as *Ozarcodina pirata* Uyeno and Barnes and *Apsidognathus* sp. were discovered at the level of 13.5 m above the base. The uppermost part with scattered fragments of similar rocks is unexposed. Thickness: 15.5 m.

Turf-covered interval with single rock clasts lithologically similar to those from member 1.

(2) Sandy limestone, thick-platty (up to 1 m), gray to blue-gray on the fresh surface and yellow on the weathered surface. The rock encloses rare interlayers (up to 15 cm) of green schistose siltstone. Thickness: 5 m.

(3) Similar limestone alternating with gray highly sandy limestone, which shows a cellular structure by weathering, and with detrital, blue-gray to dark-gray limestone containing abundant thin calcite veins. Thickness: 20 m.

Section IV. In the northeast-striking ridge, situated on the right bank of the Gromotukha River 100 m downstream from Section III, there are outcrops of the following beds (from the base upward):

(1) Layers (0.5–2.5–10 cm) of green siltstone and fine-grained sandstone alternating with rare flows (up to 15 cm) of sheared basalt and with layers (5–10 cm) of gray crystalline limestone confined to the middle part. Thickness: 70 m.

(2) Gray sandy limestone, which shows a cellular structure by weathering with a yellowish-brown network formed by more solid sandy material. A discovered fragment of Pa-element presumably belongs to *Apsidognathus* sp. Thickness: 1.5 m.

(3) Layered sandstone, yellow to light-brown, fine- to medium-grained, with isolated interlayers (up to 30 cm) of gray limestone segregation similar to the previous rock. Thickness: 2 m.

(4) Light-brown siltstone and fine-grained sandstone alternating with lenses and lensing interlayers (up to 15 cm) of limestone similar to the rock of unit 2. Thickness: 12.5 m.

(5) Limestone, as in unit 2, with fragments of *Panderodus* sp. Thickness: 3.5 m.

(6) Crystalline massive limestone of variable gray color and platty (up to 1.5 m) structure. Some levels of the bed are rich in remains of Bryozoa, Rugosa, and Crinoidea. The upper third encloses several interlayers of crinoid limestone. At the level of 1.5 m above the base, conodonts are represented by *Apsidognathus tuberculatus* and *Panderodus unicostatus*. The next level (11 m above the base) yielded *Tuxekania* sp., *Distomodus* sp., and *Panderodus unicostatus*. Thickness: 17 m.

(7) Limestone, light- to dark-gray with pink and red patches, fine-grained and massive, containing remains of Bryozoa and Rugosa. Thickness: 30 m.

ANALYSIS OF FOSSIL ASSEMBLAGES

The rich conodont assemblage that characterizes the stratotype and the parastratotype of the Yarovka Horizon and correlative beds consists of the following species: *Pterospathodus celloni, P. pennatus pennatus, P. amorphognathoides, P.* cf. *praeamorphognathoides, Distomodus staurognathoides staurognathoides, Apsidognathus tuberculatus arcticus, Aulacognathus bullatus, Johnognathus huddlej, Tuxekania* sp., *Ozarcodina* cf. *excavata, Oz. pirata, Pseudobelodella silurica, ? Oulodus fluegeli, Pseudoneothodus tricornis, Panderodus gracilis*, and *P. unicostatus*, Gen. and sp. indet 1, 2. Because of a high proportion of cosmopolitan species, this assemblage is well-correlative with coeval assemblages from Central Asia (Tuva, Mongolia), Europe, and America (Walliser, 1964; Aldridge, 1975, 1979; Sawage, 1985; Uyeno and Barnes, 1983; Armstrong, 1990).

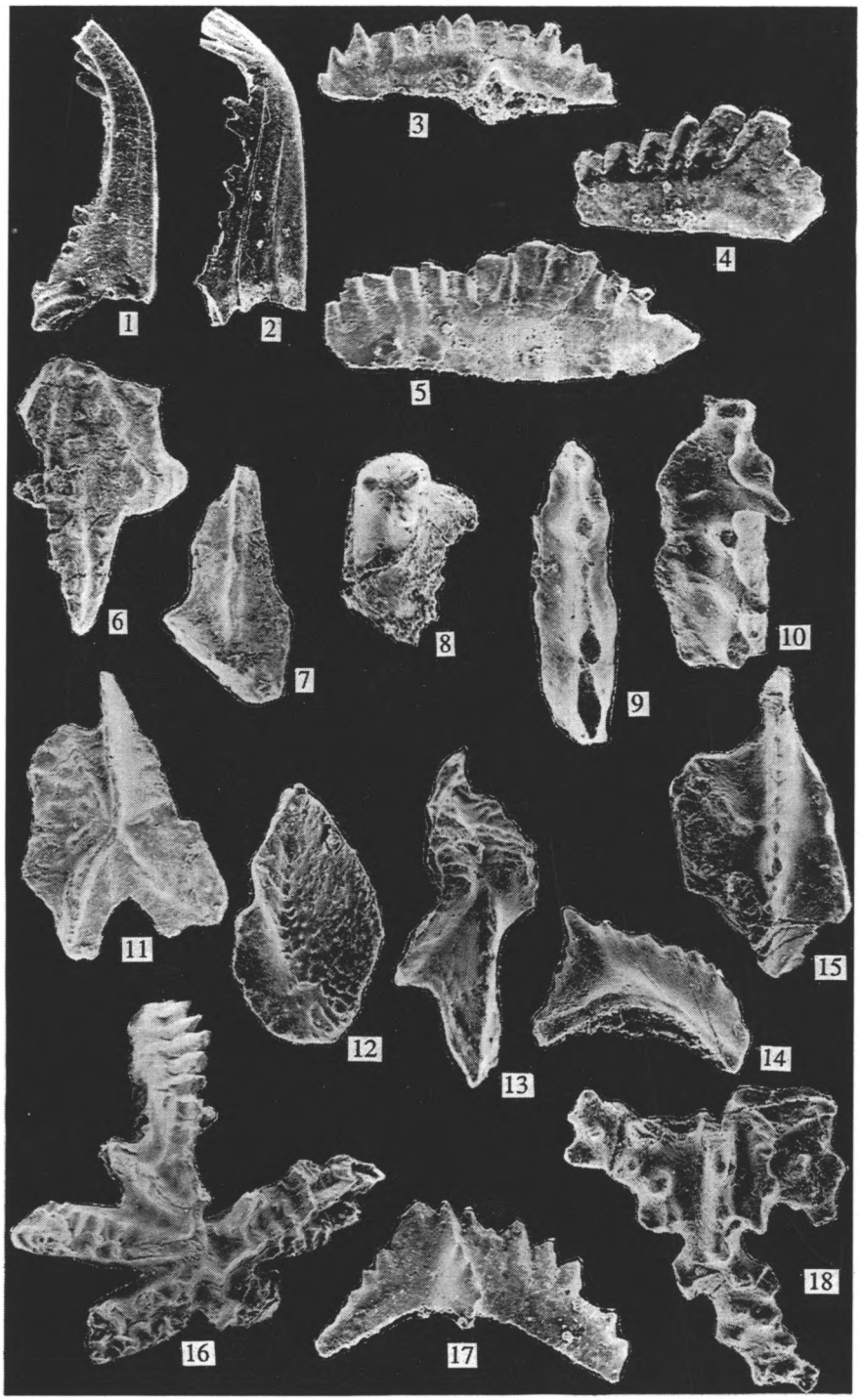
It is likely that the section encloses an upper part of the *celloni* Zone with *Pt. celloni* and *Apsidognathus tuberculatus arcticus* and a basal portion of the *amorphognathoides* Zone with *Pterospathodus amorphognathoides* and *Johnognathus huddlej*.

Thus, despite the facial difference, sections I and II with similar conodont assemblages (Fig. 3) correspond to the upper Llandoverian, which is in agreement with previous age determinations for the Yarovka Horizon (Kul'kov, 1967, 1973, 1974, 1989; Sennikov, 1976).

Species from the Altai Mountains are very similar to conodonts from southeastern Alaska (Sawage, 1985), because they include representatives of the peculiar taxa *Tuxekania* (the *amorphognathoides* Zone). Their assemblage is especially correlative to the late Llandoverian conodonts from northern Greenland (Armstrong, 1990), as some of the studied specimens are represented by *Pseudobelodella silurica, Apsidognathus tuberculatus arcticus, Ozarcodina pirata,* and *Tuxekania* sp., reported from the top of the *celloni* Zone and from the *amorphognathoides* Zone. Similar conclusions have been suggested by Kul'kov (1985) on the basis of brachiopod fauna.

Both the Yarovka Horizon of Altai and the coeval deposits of Greenland correspond to the first occurrence level of taxa that became widespread in later geological epochs. These are representatives of *Gypidula, Howellella*, and some other taxa (see plate).

The mixtite unit of Section I, which shows fragments of shallow-water carbonate deposits (including algal limestones) transposed and embedded into the



Conodonts from the Yarovka Horizon.

1, 2, *Pseudobelodella silurica* Armstrong: 1, GIN No. 1422, p-element; 2, GIN No. 1423, a-element; both from sample 90-1/4, ×60. 3, *Pterospathodus celloni* (Walliser), GIN No. 1424, Pa-element, sample 90-II/2, ×60. 4, 5, *Ozarcodina pirata* Uyeno and Barnes: 4, GIN No. 1425; 5, GIN No. 1426; both Pa-elements from sample 90-III/1, ×80. 8, *Pseudoneothodus tricornis* Drygant, GIN No. 1429, sample 90-II/2, ×35. 9, 10, *Johnognathus huddlej* Mashkova: 9, GIN No. 1430, Pa-element; 10, GIN No. 1431, Pa-element; both from sample 90-II/2, ×35. 11–14, *Apsidognathus tuberculatus arcticus* Armstrong: 11, GIN No. 1432, Pa-element; 12, GIN No. 1433, Pb-element; 13, GIN No. 1434, M-element; 14, GIN No. 1435, Sa-element; all from sample 90-II/2, ×35. 15, *Apsidognathus tuberculatus* Walliser, GIN No. 1436, Pa-element, sample 90-IV/5, ×40. 16, *Distomodus staurognathoides* (Walliser), GIN No. 1437, Pa-element, sample 90-II/2, ×35. 17, 18, *Aulacognathus bullatus* (Nicoll and Rexroad): 17, GIN No. 1438, Pb-element; 18, GIN No. 1439, Pa-element; both from sample 90-II/4, ×40.

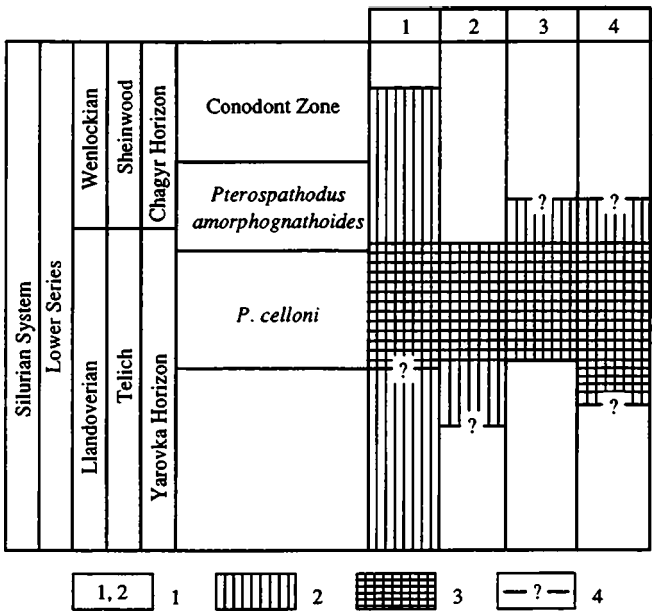


Fig. 3. Correlation of studied sections.
1, section numbers; 2, section intervals; 3, intervals with conodonts; 4, unclear position of boundaries.

siltstone matrix, proves Sennikov's idea (1976) that certain areas of the Silurian basin were favorable for the gravity sliding of sediments. Observations suggest that, during the late Llandoverian time, large masses of carbonate material were transported from a shallow setting into an adjacent depression accumulating terrigenous graptolite-bearing deposits. The occurrence of such relatively deep-water forms as *Pseudobelodella silurica* (Armstrong, 1990) in Section I is additional evidence in support of this idea. The high variability of rocks in the Yarovka Horizon and all of the above considerations point to a very irregular bottom relief in the late Llandoverian sea and to peculiar dynamics of sedimentation in this structure.

CONCLUSION

(1) The late Llandoverian conodonts of the *celloni* and *amorphognathoides* zones are distinguished for the first time in the terrigenous-carbonate deposits of the Altai Mountains as an important faunal group of the Lower Silurian Yarovka Horizon.

(2) The conodont species from Altai are most similar to coeval forms reported from southeastern Alaska and northern Greenland.

(3) Lithologic features of the studied sediments suggest that the bottom topography of the late Llandoverian sea was highly irregular.

Reviewers I.S. Barskov and A.Yu. Rozanov

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BRIEF COMMUNICATIONS

Diatom Flora from the Upper Quaternary Sediments of the Caspian Sea

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Diatoms are still insufficiently used for the stratigraphic subdivision of the Caspian Sea sediments. There are only isolated publications, and the number of studied sections is insignificant for a major intracontinental basin as large as the Caspian Sea (Zhakovshchikova, 1974; Proshkina-Lavrenko and Makarova, 1968; Sorokin *et al.*, 1984). The stratigraphic interpretation of the available data is not clear because of the variable volume of the stratigraphic horizons, which were established by different researchers. Therefore, this greatly effects the estimation of the thickness of these units. In this situation, it is diatoms that make it possible to stratify the Caspian deep-sea deposits, although they lack remains of mollusks. The latter are the most reliable instrument for shelf stratigraphy. Aside from solving the applied problem, the present work considers in detail the composition of diatom assemblages in different facies zones of the central and southern Caspian Sea (the shelf, continental slope, and deep-sea depressions), which allows definite paleogeographic conclusions to be drawn.

This article is based on nine cores of Upper Quaternary deposits (Fig. 1). The cores were normally sampled at 20-cm-intervals in all lithologic beds; some of them are barren of diatoms.

The studied sediments differ in composition. Thus, in the central Caspian Sea (Stations 15 and 67), the following horizons were detected from top to bottom. The first horizon consists of gray-green and green clayey muds enriched in organic matter (up to a distinct sapropelic interlayer) and diatoms, which form thin layers. These sediments rest upon gray and pinkish-gray clayey mud with lumps and aggregates of hydrotroilite. The bed is underlain by compact light brown clayey mud with stripes of hydrotroilite, which transforms into compact blue mud. It is worth noting that the gray mud is enriched in coarsely grained sandy and silty material of turbidite origin. Sometimes, even distinct layers are observed. A similar section is described at the shallower Station 66. This section is distinct in the somewhat greater thickness of the gray mud. At Stations 56 and 58, the upper part of the sections demonstrate a bed of gray and green diatomaceous and clayey mud with thinly stratified zones. The bed is underlain by gray clayey mud. The diatomaceous ooze of these and the above-mentioned columns has a high content of amorphous silica (12%). This is an order of magnitude greater than that in the underlying clayey muds. At Station 59, the section is represented by predominantly gray silty and clayey mud with silty lenses and an impoverished molluscan association. In the western shelf of the central Caspian, core 35 recovered a twofold section. Its upper part is formed by silty and clayey mud with a high content of molluscan shells; the lower part is composed of thinner mud with less frequent mollusks. In the outer shelf (Station 36), the section is capped by a coquina layer. It is underlain by clayey mud with interlayers of silts and local enrichment in molluscan shells. The bed is underlain by brown clayey mud with thin silty layers and rare shells. In the central part of the southern Caspian, a thin bed of greenish-gray diluted clayey mud is underlain by brownish-gray clayey mud with numerous interlayers of hydrotroilite followed by homogenous dark gray to almost black clayey mud. In the lowermost part of the section, the mud acquires a light gray color (Fig. 2).

Before presenting our material, a brief discussion of the modern diatom flora of the Caspian Sea is needed. Diatoms of the Caspian Sea are distinct from the diatom algae of basins with normal salinity in their relatively low content of marine elements, the diversity of the brackish-water species, and the presence of freshwater species (primarily in the northern Caspian Sea). As a whole, the assemblage has the following characteristics: a considerable content of some marine species, such as *Rhizosolenia calcar-avis* Schul., *Coscinodiscus radiatus* Ehr., and *Thalassionema nitzschioides* Grun.; a high diversity of marine brackish-water species of the genus *Chaetoceros* Ehr. (*Ch. wighamii* Brig., *Ch. simplex* Ost., and others), *Coscinodiscus jonesianus* (Grev.) Ost., *Actinocyclus ehrenbergii* Ralfs., *Synedra tabulata* (Ag.) Kütz., and others; and brackish-water species *Thalassiosira caspica* Makar., *Th. incerta* Makar., *Th. variabilis* Makar. (endemic forms), *Cocco-*

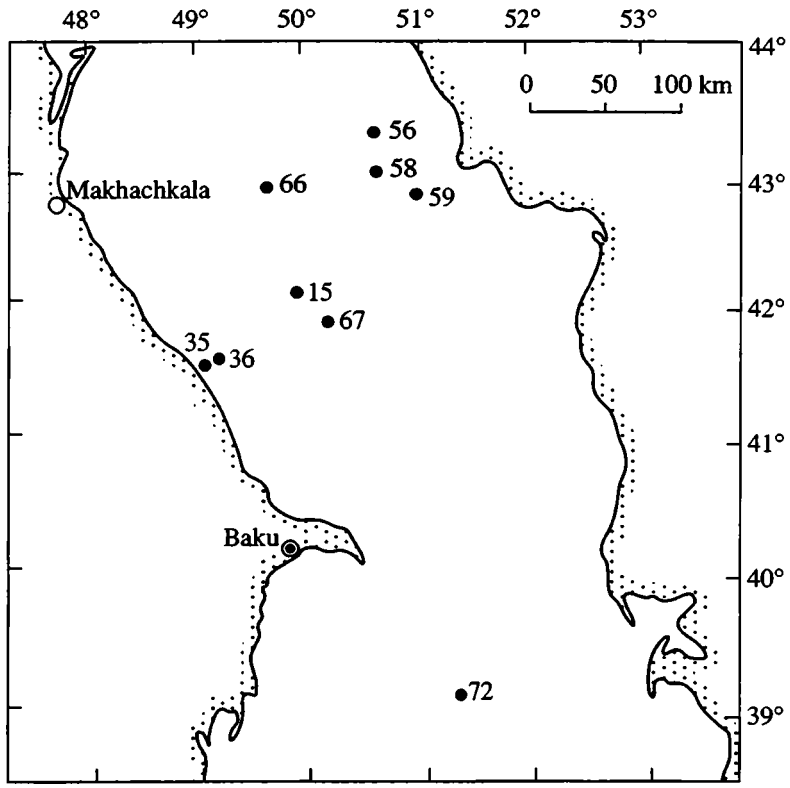

Fig. 1. Location scheme of the studied cores. Figures refers to the numbers of the stations.

neis scutellum Ehr., and others. Freshwater and freshwater–brackish-water species of the genera *Aulacosira* Thev., *Stephanodiscus* Ehr., and *Diatoma elongatum* (Lyngb.) Ag. are more common in the northern Caspian Sea (Proshkina-Lavrenko and Makarova, 1968). The diatom assemblage of the northern Caspian Sea differs from the assemblages of the central and southern areas. It has typical estuarine features, whereas the two other assemblages are marine ones, though they are strongly impoverished in marine elements and contain a high number of marine brackish-water species.

The study of the species and ecological composition of the fossil diatom flora resulted in the establishment of the two assemblages, which sequentially replace each other along the sedimentary section.

The upper assemblage (I) represents greenish-gray muds with an admixture of sapropel in the deeper areas and silty and clayey muds with molluscan shells in the shelf. The assemblage is characterized by the domination of marine planktonic species; *Coscinodiscus radiatus* Ehr. plays the leading role. Also present are marine *C. gigas* Ehr., *Rhizosolenia calcar-avis* Schul., *Rh. alata* (Ehr.) Brig., and *Thalassiosira excentrica* (Ehr.) Cl. and marine brackish-water species *Actinocyclus ehrenbergii* Ralfs., *Cyclotella caspia* Grun., and *C. striata* (Kütz.) Grun., and *Coscinodiscus jonesianus* (Grev.) Ost. Benthic marine and freshwater species are rare and generally typical of the northern Caspian Sea. The composition of the established assemblage shows some changes with regard to facies zones and marine areas (table). In this respect, the most significant changes are found in the cores from the western shelf of the central Caspian Sea. For example, the core from Station 35 shows a relative increase in marine and freshwater benthic forms. This is an indication of shallow conditions and the influence of the Samur River. The lower part of the core demonstrates an increasing role of brackish-water and freshwater species. At the margin of the shelf (Station 36), the section also shows similar changes in the composition of marine planktonic forms and the appearance of considerable numbers of benthic marine species and planktonic and, particularly, benthic freshwater species. This may point to shallower conditions and a somewhat decreased salinity.

Thus, the described assemblage is dominated by planktonic marine species. Marine brackish-water species play a much smaller role. The species composition of the diatom flora also indicates the predominance of ubiquists, the subordinate role of species of temperate latitudes, and rare occurrences of endemic species (*Chaetoceros diversicurvatus* V. Goor and *Stephanodiscus binderanus* (Kütz.) Krieg., Station 67). Noteworthy are finds of cold-water species *Chaetoceros subsecundus* (Grun.) Hust. and *Ch. holsaticus* Sch. in the western shelf of the central Caspian Sea (Station 36).

Table. (Contd.)

II	Depression (B)	Eastern slope (A)		Depression (A)
	Station 72 (depth 970 m)	Stations 59 (depth 190 m)	Stations 56, 58, 66	Stations 15, 67
1	*Rhizosolenia calcar-avis* Schul. *Coscinodiscus radiatus* Ehr. *Thalassiosira oestrupii* (Ost.) Hasle *Th. decipiens* (Grun.) Jorg.	*Thalassiosira excentrica* (Ehr.) *Actinocyclus divisus* Kiss. *Diploneis subcincta* (A.S.) Cl. *Coscinodiscus radiatus* Ehr.	*Thalassiosira excentrica* (Ehr.) Cl. *Rhizosolenia calcar-avis* Schul. *Rh. alata* (Ehr.) Brig. *Coscinodiscus radiatus* Ehr. *Actinocyclus divisus* Kiss.	*Rhizosolenia calcar-avis* Schul. *Rh. alata* (Ehr.) Brig. *Rh. fragilissima* Berg. *Coscinodiscus radiatus* Ehr. *C. perforatus* Ehr. *Thalassiosira oestrupii* (Ost.) Hasle
2	*Actinocyclus ehrenbergii* Ralfs *Coscinodiscus jonesianus* (Grev.) Ost.	*Actinocyclus ehrenbergii* Ralfs *Diploneis interrupta* (Kütz.) Cl.	*Actinocyclus ehrenbergii* Ralfs. *Cyclotella caspia* Grun.	*Actinocyclus ehrenbergii* Ralfs.
3	*Thalassiosira incerta* Makar. *Th. caspica Makar.* *Th. coronifera* Pr.-Lavr.			*Thalassiosira caspica* Makar. *Actinocyclus paradoxus* Makar. *Sceletonema subsalsum* (Cl.) Beth. *Cocconeis scutellum* Ehr. *Diploneis smithii* (Breb.) Cl. *Nitzschia hungarica* Grun.
4	*Stephanodiscus astraea* (Ehr.) Grun. *Diploneis ovalis* (Hilse) Cl. *Cocconeis placentula* Ehr.	*Cyclotella kützingiana* Thw.	*Cyclotella kutzingiana* Thw. *Stephanodiscus astraea* (Ehr.) Grun. *Cyclotella operculata* (Ag.) Kütz. *Stephanodiscus binderanus* (Kütz.) Krieg.	*Nitzschia tryblionella* Hant. *Diploneis ovalis* (Hilse) Cl.
5	*Synedra ulna* (Nitzsch.) Ehr. *Aulacosira granulata* (Ehr.) Sim.	*Navicula reinhardtii* Grun. *Synedra ulna* (Nitzsch.) Ehr. *Epithemia zebra* (Ehr.) Kütz.	*Cyclotella comta* (Ehr.) Kütz. *Paralia arenaria* (Meor.) Moiss. *Hantzschia amphioxys* (Ehr.) Grun. *Fragilaria brevistriata* Grun.	*Synedra ulna* (Nitzsch.) Ehr. *Cymbella aspera* (Ehr.) Cl. *C. helvetica* Kütz. *Fragilaria construens* (Ehr.) Grun. *Navicula amphibola* Cl.

Abundance: — — — single; ——— rare; ═══ common. Diatom species: *1*, marine; *2*, marine brackish-water; *3*, brackish-water; *4*, freshwater–brackish-water; *5*, freshwater.

assemblage, the lower one includes mainly brackish-water forms, which indicates lower salinity during the formation of the enclosing sediments. The presence of freshwater benthic diatoms in deep-sea sediments (Stations 67 and 72) resulted from their removal from the shelf by turbidity flows. The assemblage is dominated by species of temperate latitudes. The presence of moderately warm-water species is remarkable. The group includes numerous endemic forms, such as *Actinocyclus paradoxus* Makar., *Thalassiosira caspica* Makar., and *Thalassiosira incerta* Makar. Deposits at Station 59 yielded three cold-water species. One of them, *Diploneis interrupta* (Kütz.) Cl., was marked as a "mass occurrence."

A joint analysis of the diatom flora and the lithology together with the palynological data (Sorokin and Chernysheva, 1983) from some of the cores, which are described in the present article, make it possible to subdivide the studied deposits into two horizons.

The deposits of the Novocaspian Horizon are enriched in valves of diatoms and organic matter. The horizon contains the upper assemblage of diatom flora (I) and is distinct in the domination of marine, mainly cosmopolitan species. The spore and pollen assemblages have a maximum content of arboreal pollen, including those of broad-leaved forms. All of this indicates that sedimentation occurred during the Holocene (beginning in the Atlantic period) under a rise in the sea level and relatively high salinity. This is confirmed by the presence of marine mollusks in shallow-water deposits. The thickness of the Novocaspian Horizon in the studied sections outside the shelf zone varies from 50 to 160 cm. At some stations (59 and 72), it does not exceed 20 cm, which is apparently due to erosion. Brownish and gray muds were recovered at Station 36 120 cm below the sea floor. These deposits yielded a somewhat different diatom complex. The palynological assemblage of the same deposits demonstrates an abrupt decrease in arboreal pollen and an increase in its reworked forms. This indicates a sea-level drop under climatic aridization and, therefore, the transitional nature of the deposits.

The Upper Khvalynian Horizon is represented by gray and brown muds. It contains the lower diatom assemblage (II) with dominant marine brackish-water species. Freshwater–brackish-water species are less common, and marine species play an insignificant role. A considerable part of the assemblage is represented by boreal forms. Cold-water species were detected in the northeastern part of the central Caspian Sea. The deep central part of the sea shows a diversity of endemic forms. The lower part of the section and, particularly, the brown muds almost completely lack diatoms. In the pollen assemblages, xerophytic herbs play the leading role; in the upper part of the horizon, however, an increase in arboreal pollen is observable. The presented data suggest that sedimentation occurred in an arid climate with low salinity; this was accompanied by a rise in sea level from the minimum values of the last (Ostashkov) glaciation to the higher position in the pre-Holocene time.

CONCLUSION

(1) Two assemblages of diatom flora are established in the section of the upper Quaternary deposits of the Caspian Sea. The lower one corresponds to the Upper Khvalynian Horizon, and the upper one correlates with the Novocaspian Horizon. The diatom-based age subdivision of the sections matches well with the results of a study of the palynology and mollusks (Sorokin *et al.*, 1984).

(2) The compositions of the diatom assemblages and their changes over time indicate that the sedimentation of the Upper Kvalynian deposits occurred under conditions of low salinity, low sea level, and an arid climate. This time corresponds to the late stage of the Valdai Glaciation. The Novocaspian beds were deposited under conditions of increased salinity and high sea level during the Holocene transgression.

Reviewer M.A. Akhmet'ev

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